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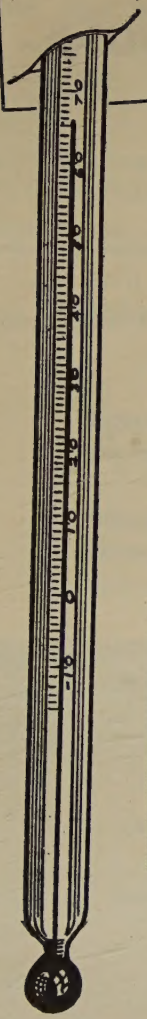

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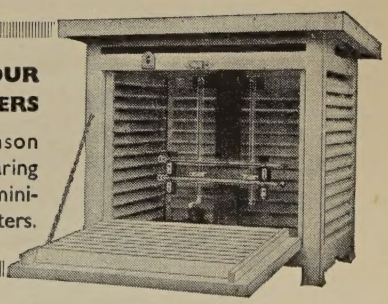

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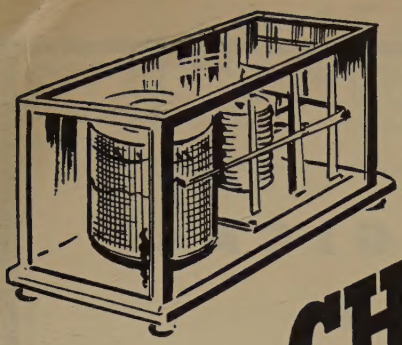

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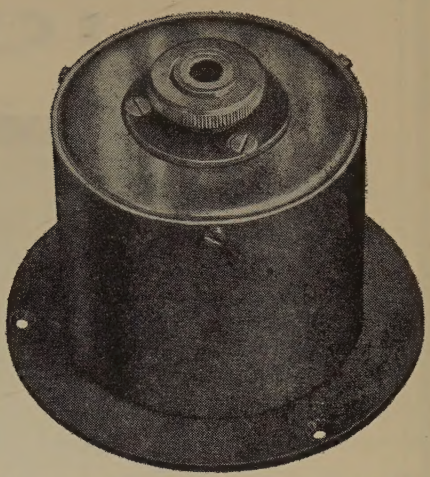

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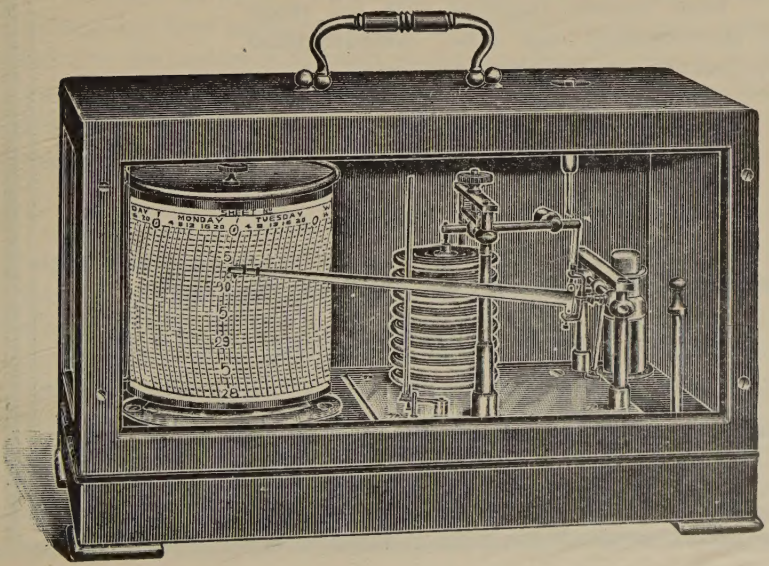

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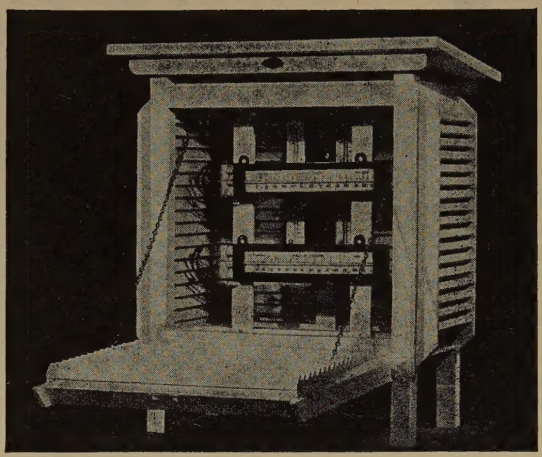

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The forms of ice clouds: II

By F. H. LUDLAM
Imperial College, London

(Manuscript received 24 January 1956, in revised form 16 April 1956)

SUMMARY

The role of convection (released by the latent heat of condensation when crystals fall into ice-supersaturated air) in the regeneration of fallstreak cirrus forms is discussed and illustrated.

1. INTRODUCTION

In a previous paper (Ludlam 1948, referred to subsequently as I) I explained that the most typical form of isolated ice clouds, *cirrus uncinus*, can be regarded as a trail of precipitation. The crystalline particles in such a trail acquire sizes and fall-speeds characteristically much greater than those of the particles in a recently formed droplet cloud. This is primarily because they form upon special nuclei whose concentration is very much smaller, and because, forming at a high ice supersaturation, the crystals continue to grow as they settle through a layer several hundred metres deep in which the air remains supersaturated with respect to ice.

Recently Douglas (1955), following work with Marshall, has emphasized that very similar, though sometimes larger, trails are discovered by radar examination to be common features of cloud systems giving extensive snow at the ground. These workers have shown that the snow trails are descending from generating regions which move with the wind, and that about half of the continuous snow which reaches the ground in Canada is produced by the procession overhead of several such rather small sources (see Ludlam 1956). Each trail is spread out by wind shear, and successive trails may merge before their crystals reach the ground.

To explain the apparently continuous generation of snow crystals in localized regions, the Canadians have suggested that crystal growth in the trails warms the air there sufficiently (by the release of the heat of condensation) to cause it to rise and renew crystal formation in the region adjacent to the head of the trail. They suggest that this process, and other properties of the snow trails, might be illustrated by the study of fair-weather cirrus. The following paragraphs show that this is indeed so.

2. THE RELATION BETWEEN CIRRUS FALLSTREAKS AND SHRED-CLOUDS

In the previous paper (I) the forms of cirrus fallstreaks were shown to be quite similar to those of the crystal trails which are often seen to fall from isolated mother-clouds in the altocumulus levels, composed of supercooled droplets. Cirrus fallstreaks often have no mother-cloud, and are apparently formed at humidities rather less than 100 per cent with respect to liquid water. Nevertheless it was mentioned that shreds of non-fibrous cloud often occur at the heads of cirrus fallstreaks (I, Fig. 1 (a), Plate II), and it was implied that these represent a kind of mother-cloud. Since that time, however, I have repeatedly observed that these shreds do not form until the fallstreak is well developed, and that often they are separated from the top of the fallstreak by a distinct gap. Figs. 1 (a) and 1 (b) illustrate such an occurrence. The distance between the top

of the fallstreak and the shred-clouds usually amounts to a few hundred metres. When the thick part of the fallstreak is almost vertical, as in the figure, the shreds appear directly above the top of the fallstreak, but when the fallstreak is inclined by a wind-shear the shreds are set aside from its top in the direction of the shear : in fact, they lie above the lower parts of the fallstreak. These observations show that although the fallstreak and the shred-cloud are intimately related, it is the fallstreak which is the mother-cloud. Fig. 2 shows another occasion on which the formation of shred-clouds occurred above a rather dense mass of fallstreaks.

Apparently cloud-forming updraughts must exist above the tops of almost vertical fallstreaks, and above the length of strongly inclined fallstreaks. The only obvious cause of such localized updraughts is the warming of the air in the fallstreaks themselves. Sometimes the fallstreaks beneath the shred-clouds become distorted in a way which clearly indicates the rise of air out of them.

3. The liberation of latent heat in fallstreaks

When ice crystals settle into air with a high ice-supersaturation they grow rapidly by condensation. Their surface temperature rises above the air temperature, and the heat of condensation is communicated to the air. The maximum release of heat occurs when sufficient vapour is condensed to bring the air to a state of ice saturation, and can readily be estimated on the assumption that the heating is isobaric. In general, then, if ρ_a is the air density and ρ_1, ρ_2 the initial and final vapour densities, the increment ΔT in the air temperature T is given by

$$\rho_a c_p \Delta T = L_s (\rho_1 - \rho_2) \quad . \quad . \quad . \quad . \quad (1)$$

where c_p is the specific heat of air at constant pressure (0·24 cal deg^{-1} C g^{-1}) and L_s is the latent heat of sublimation (680 cal g^{-1}).

Since

$$\rho_a = p/2870\ T,$$

where p is the air pressure in mb and

$$\rho_v = e_v/4620\ T,$$

where ρ_v and e_v are the density and pressure of water vapour, the relation (1) can be written

$$0{\cdot}24 \times 4620\, p\, \Delta T = 2870 \times 680\, (e_1 - e_2)$$

or

$$\Delta T = 1760\, (e_1 - e_2)/p \quad . \quad . \quad . \quad (2)$$

At a given pressure level the maximum value $\Delta T'$ of ΔT corresponds to a value of e_1 representing saturation over liquid water at temperature T, e_w, and the value of e_2 representing saturation over ice at temperature $(T + \Delta T')$, e_i'. The latter is related to the saturation vapour pressure e_i over ice at the temperature T by the Clausius-Clapeyron equation

$$e_i' - e_i = 6140\, e_i\, \Delta T'/T^2$$

Then

$$\Delta T' = 1760\, (e_w - e_i - 6140\, e_i\, \Delta T'/T^2)/p$$

or

$$\Delta T' = 1760\, (e_w - e_i)/(p + 1{\cdot}08 \times 10^7\, e_i/T^2) \quad . \quad . \quad (3)$$

Table 1 gives values of $\Delta T'$ as a function of temperature, for pressures appropriate to conditions near the British Isles in summer air masses or maritime tropical air. The

Figure 1. Formation of shred-clouds above *cirrus uncinus*, 2000 hr, 6 July 1951, Dunstable. The interval between the two pictures is 8 min.

Figure 2. Formation of cumuliform shred-clouds above cirrus fallstreak mass, 1005 hr-1055 hr, 2 Sept. 1949, Dunstable. The clouds are moving away from the observer, towards the NE., but the degree of enlargement in each picture has been adjusted to keep the entire cloud mass approximately the same size. The interval between successive pictures is about 4 min, increasing to 10 min and 25 min between the next-to-last and the last pairs.

values of e_w and e_i are taken from the formulae adopted by the Aerological Commission of the O.M.I. (1951); those of e_i are regarded as thoroughly acceptable, but those of e_w are extrapolations from higher temperatures, not supported by any experimental data below — 15°C.

TABLE 1. VARIATION WITH TEMPERATURE OF $\Delta T'$, THE MAXIMUM TEMPERATURE RISE DUE TO RELEASE OF HEAT OF CONDENSATION WHEN THE HUMIDITY IS REDUCED ISOBARICALLY TO A STATE OF SATURATION OVER ICE

Temperature °C	$(e_w - e_i)$ mb	Average p, mb	$\Delta T'$ °C	γ °C/km	d m	z m
— 10	0·27	515	0·51	6·5	150	140
— 20	0·22	420	0·65	7·5	260	250
— 30	0·13	350	0·54	8·4	340	350
— 40	0·061	300	0·33	(8·7)	250	420
— 50	0·024	250	0·16	(9·0)	160	460
— 60	0·008	220	0·06	(9·0)	60	490
— 70	0·002	180	0·02	(9·0)	20	500

4. THE FORMATION OF SHRED-CLOUDS ABOVE FALLSTREAKS

Table 1 also includes typical values of the lapse-rate γ. In the medium levels in maritime tropical air the mean lapse rate is very close to the saturated adiabatic (with respect to liquid water), according to the data of Belasco (1952), and at temperatures down to — 30°C the values in the table are those of the saturated adiabatic lapse rate. According to a diagram published by Murgatroyd, Goldsmith and Hollings (1955) representing the results of a number of aircraft soundings in southern England, the mean lapse rate in the 2 to 3 km below the tropopause is only about 6°C/km, but this seems rather small, and, for temperatures of — 40°C and below, the values in the table are representative of those occurring on the occasions of the observations of cirrus listed in Table 2.

From these values it is possible to derive the distance which convective bubbles may rise, after acquiring a temperature excess $\Delta T'$, before again reaching temperature equilibrium with the environment. It is first necessary to consider whether the bubbles carry crystals. The sizes of fallstreaks are rather variable, but usually they are seen to have a horizontal extent of a few hundred metres, and it may be supposed that the bubbles of warmed air formed in the trails have about this diameter. A bubble of this size, with a temperature excess of 0·1 to 0·4°C, has a limiting speed of between 1 and 2 m/sec (Scorer and Ludlam 1953). In a stable atmosphere the temperature excess decreases as the bubble rises, and as it becomes diluted by mixing with outside air, so that rising speeds are not likely to exceed 1 m/sec. The large ice crystals which are found in well-developed fallstreaks have fall speeds of $\frac{1}{2}$ m/sec or more (I). Unless they are unusually small, they must be left behind by the bubbles emerging from a crystal trail, and therefore as the bubbles rise they cool at the dry adiabatic rate. Under these conditions the maximum vertical ascent d to the new equilibrium level is that shown in Table 1. The last column of the table shows the distance z which air originally saturated with respect to ice must rise if it is to become saturated with respect to liquid water. At temperatures down to nearly — 40°C the values in the last two columns are about equal, and therefore it is to be expected that the bubbles rising from an inclined fallstreak produce fresh clouds. Since the bubbles may 'overshoot' their equilibrium level by about a radius (one or two hundred metres), and because the clear-air lapse rate occasionally exceeds the quoted values, we may expect shred-clouds to form in this way commonly at temperatures down to about — 40°C, but rarely if at all at temperatures below — 50°C.

It should be noted that during the first part of the rise of the bubbles, while they are within the crystal trail, the process is neither isobaric nor adiabatic. When this part is a small fraction of the total rise the actual process will have results not very different from those estimated by assuming isobaric heating succeeded by dry-adiabatic cooling; otherwise the total possible rise may be increased. A thorough examination of the actual process would require the specification of the thickness of the crystal trail, the size and fall-speed of its crystals, the rate at which the originally supersaturated air is reduced to the saturated state, and the external lapse rate, and is not attempted here.

TABLE 2. OBSERVATIONS OF CIRRUS AT DUNSTABLE (1948-1953) AND AT ÖSTERSUND (1954).

Date	Cloud forms F = fallstreaks, S = shred-clouds, C = cumuliformis.	Pressure (mb) and temp. (°C)		Pressure range over which lapse rate observed, mb	Lapse rate (°C/km)	
		Cloud top	Cloud base		Observed value	Saturated adiabatic at cloud base
5 Jan. 1948	Dense F; some S and C	370 − 43	420 − 45	350-450	8·5	8·6
10 Oct. 1948	Well developed C; sparse F	250 − 47	380 − 23	250-350	8·1	7·7
27 Oct. 1948	Dense F; S; some C	425 − 37	500 − 28	425-500	7·5	8·7
13 May 1949	Dense F in lines; no trace of S	240 − 60	280 − 50	250-300	9·1	9·7
6 June 1949	Dense F; S; well developed C; a globular tower seen to grow from F and acquire and penetrate a pileus.	330 − 40	450 − 23	330-450	8·4	8·0
27 June 1949	Dense F and S; well developed C	230 − 50	300 − 39	250-300	8·7	9·1
2 July 1949	Moderate F and C	350 − 31	425 − 23	350-450	7·5	7·9
3 July 1949	Dense F; S	330 − 37	380 − 27	330-400	8·3	8·3
4 July 1949	Large acu. C with dense mamm. F	470 − 19	600 − 5	500-600	8·1	5·8
11 July 1949	Bands of dense F and C	250 − 50	330 − 34	250-300	8·7	8·9
26 July 1949	Bands of F with some C; formed from acu.	480 − 15	550 − 7	500-550	6·4	6·9
4 Aug. 1949	S formation seen above sparse F	250 − 47	335 − 33	250-300	8·1	8·8
2 Sept. 1949	Moderate F; S	335 − 40	370 − 35	300-400	8·7	9·0
4 Mar. 1950	Moderate F; C	275 − 50	370 − 33	250-350	8·5	8·9
14 Aug. 1952	Dense F; some S and C	320 − 43	450 − 23	350-450	8·7	8·0
17 Aug. 1953	Dense F; S and C	300 − 47	350 − 37	300-350	8·3	9·1
12 Sept. 1954	Dense F; some S	330 − 43	400 − 35	350-400	7·9	9·1
30 July 1954	Moderate F; S and C	340 − 41	400 − 32	350-400	7·5	8·9
24 July 1954	Dense F, mamm; S and C	290 − 45	340 − 37	300-340	7·4	9·1
	Thin F; no S	250 − 53	—	250-300	7·7	—
20 July 1954	Moderate F; S	310 − 44	350 − 37	308-350	9·0	9·1

5. Observations of cirrus fallstreaks and shred-clouds

Table 2 is a list of occasions on which cirrus with fallstreaks and shred-clouds has been observed at Dunstable, and, in 1954, at Östersund in Sweden. These forms of cirrus are very common, and the listed occasions are only those on which careful observations happen to have been taken, and on which a reliable estimate of the height could be made. At Dunstable the height was reported by aircraft on about half of the occasions; on the remainder the height was found by relating the cloud motion (accurately measured with a theodolite) and the fallstreak directions to the winds and the wind shears observed at the neighbouring radiosonde stations. At Östersund the cloud heights were measured by the two-theodolite method. The table also shows the lapse rates at the cirrus levels (obtained from radio-sonde records), and the saturated adiabatic lapse rate (with respect to ice) at the level of the lower limit of the cirrus.

On two occasions, 13 May 1949, and 24 July 1954, when it was especially noted that shred-clouds did not occur with the fallstreaks, the cirrus-top temperatures are low (— 60°C and — 53°C). On all occasions when shred-clouds were present these temperatures are higher, and several hundred metres below the cloud tops, where the air in the shred-clouds may be expected to begin its rise, the temperature is never lower than — 45°C. On 24 July 1954, it was noticed that fallstreaks with pronounced shred-clouds and cumuliform heads were at a lower level and higher temperature (above — 44°C) than the thinner fallstreaks without shred-clouds. These observations support the theory of shred-cloud formation.

On a few occasions (as for example, 2 September 1949, illustrated in Fig. 2), when the fallstreaks were strongly inclined and the shred-clouds clearly separated from them, the vertical separation was estimated, and can be compared with values of z derived from Table 1, with results shown in Table 3.

TABLE 3. Vertical separation of shred-clouds from fallstreaks

Date	Separation (m)	Temperature at fallstreak level (°C)	z (m)
27 June 1949	820	— 39	400
2 July 1949	300	— 23	280
26 July 1949	250	— 7	100
4 Aug. 1949	450	— 33	380
2 Sept. 1949	350	— 35	390
	320		
	500		
	330		
17 Aug. 1953	250	— 37	400
24 July 1954	300	— 37	400

There is reasonable agreement between the observed and the theoretical values, considering the difficulty of estimating the separation of the shreds from the rather diffuse-edged fallstreaks, and of estimating the fallstreak temperatures.

6. The composition of shred-clouds

As the shred-clouds are supposedly formed when vapour in air rising from the fallstreaks reaches or closely approaches a state of saturation with respect to liquid water, it is likely that at least for a short while after their formation the shred-clouds contain droplets. It is therefore not surprising that iridescent colours can often be seen in the

clouds when they are near the sun (the observations are best made by observing reflections from a black mirror through polaroid spectacles). The colours are usually rather faint, predominantly of brownish-red or pink, but sometimes a proportion of shreds, presumably the most recently formed, show brilliant tinges of red, green and blue when about 2° from the sun. Occasionally pale iridescence can be seen as far as 8° from the sun, and rarely, in more extensive patches of cloud, I have seen a complete double corona, with an inner dull-red ring at rather less than 2° from the sun, followed by rings of blue-white and faint green, and a purple-red ring at 3° from the sun. There is usually no trace of halo phenomena in the shred-clouds, and iridescent colours may be seen at times of up to a few minutes after their formation, suggesting that the droplets are rather persistent or continually renewed. Frequently, also, shred-clouds are seen to evaporate completely within a few minutes of their formation, suggesting that on these occasions they contain no appreciable concentration of ice crystals. If such shred clouds, and those which show rather persistent iridescence, are formed at temperatures several degrees above — 40°C, then these properties are understandable. I have no reliable evidence of iridescence occurring in shred-clouds at temperatures below — 40°C.

The size and hence the concentration of the droplets in shred-clouds may be estimated from the fact that the second red ring of the corona is usually at an angular distance of about 3° from the sun. Substitution of this value in the well-known relation between angular distance θ (in degrees) and particle diameter d (in microns) (see Minnaert 1940) gives

$$d = 73/\theta \simeq 24\ \mu$$

The shred-clouds in which iridescence is seen are usually rather shallow, having a thickness of one or two hundred metres. The concentration of water condensed during a rise of 200 m at a temperature of — 35°C is about 0·045 g/m^3, and if this is shared amongst droplets of radius 12 μ their concentration is about 7/cm^3. This is of the same order as the total concentration of nuclei at these levels.

7. The regeneration of fallstreaks

If a crystal cloud is formed by the freezing of droplets having a concentration of the order 10/cm^3, it is unlikely that all of the crystals can reach the size of those which occur in fallstreaks. Weickmann (1947) found these to be hollow prisms with lengths of a few hundred microns, dimensions which must be attained if the crystals are to have the observed fall-speeds of about $\frac{1}{2}$ m/sec. Even if the crystals extract vapour over a fall path of several hundred metres through air at a high ice-supersaturation, they cannot be expected to reach such sizes unless their concentration is smaller by about two powers of ten (aufm Kampe 1952, has considered this matter in some detail).

It is therefore understandable that usually those shred-clouds which, by the freezing of most or all of their droplets, are quickly transformed into pure ice-clouds yield only faint fallstreaks. Dense fallstreaks may form, however, if the bubbles rising out of an original fallstreak produce shred-clouds at a temperature at which only a small proportion of droplets freeze (as usually happens in the range — 12°C to — 30°C). They may also form at lower temperatures if the bubbles rise too small a distance for the vapour pressure to reach saturation over liquid water, but sufficient for it obtain a high super-saturation over ice and allow the activation of rare, efficient crystal nuclei. In either of these circumstances conditions are favourable for the continual regeneration of an original fallstreak.

8. The formation of fallstreaks

An original fallstreak cloud develops in air generally ice-supersaturated throughout a deep layer, when sparse crystals form in a region a few hundred metres across. In the upper troposphere the most general cause of cloud formation is the slow, large-scale ascent of air, and the question arises why the first clouds to appear should not always have the form of a widespread shallow layer, instead of occurring as scattered fallstreaks. Probably wave-like disturbances and irregularities in the water-vapour distribution are responsible for their formation. In the peaks of orographic waves, which are practically always present overland, and over the oceans in travelling wave disturbances of other kinds, the humidity eventually reaches a value at which the rare, efficient crystal nuclei can be activated, some time before saturation with respect to liquid water is attained.

Because the rate-of-rise of the humidity is very slow, corresponding to ascending speeds of only a few cm/sec, these rare crystal nuclei have a good opportunity to be effective before other, more numerous nuclei are called upon. On the other hand, during the ascent of air at speeds of about 1 m/sec, such as occur during the formation of shred-clouds, the growth of crystals upon the rare nuclei activated at high ice-supersaturation is unable to prevent the humidity from continuing to rise and attaining liquid water saturation about a minute later. Practically all the condensation nuclei present are then activated, and at low temperatures the droplets which are formed soon freeze to produce crystal concentrations unfavourably high for fallstreak formation.

When the wave disturbances over particular ground features are very pronounced at high levels, the speed of ascent towards the crests exceeds 1 m/sec (Ludlam 1952), and if condensation occurs in the waves streamers of ice cloud are formed, notably with poorly developed fallstreaks, extending great distances downwind. Only when the wave-motions are relatively weak are fallstreaks likely, and then variations in the wave amplitude and irregularities in the humidity distribution ensure that crystal formation in stationary waves is intermittent, so that the fallstreaks are formed in isolated groups rather than any definite pattern obviously related to the topography.

9. Shred-cloud formation in unstable atmospheres

Frequently shred-clouds grow upwards for a few hundred metres in cumuliform shapes which demonstrate their convective nature. Table 2 shows that these cumuliform growths often occur when the lapse rate is generally *less* than the saturated adiabatic, even at the level of the lowermost fallstreaks. This clearly indicates that the air composing the shred-clouds has in some manner acquired a temperature excess.

When the lapse rate in the clear air reaches or exceeds the saturated adiabatic, tall cumuliform growths are produced, and fallstreak regeneration then leads to the appearance of a striking form of dense cirrus in which the individual cloud resembles a small anvil. The cumuliform tops continually form over the fallstreaks at the up-shear end of the cloud, and flatten and spread forward towards the other end, producing a characteristic wedge-shaped cloud mass with no clear distinction between the upper parts and the fallstreaks. The sharply-defined edge of the cumuliform tops is not evidence of a droplet-composition, but only of an active convection process. Equally, the fibrous nature of fallstreaks is not in itself evidence of a crystal-composition, as often believed, but only of the presence of trails of precipitation-elements, particles large enough to have fall-speeds of the order of 1 m/sec. (The usual association of a cumuliform appearance with a droplet composition, and a fibrous appearance with a crystal-composition, has led to a misconception of the significance of the transformation of tall cumulus tops into fibrous clouds which is perpetuated by the unfortunate description ' glaciation.')

The lower ends of dense fallstreaks often show a mammatus formation, due to the chilling of the air in which the crystals evaporate (I). At least a faint mottling due to this process can be seen in practically all fallstreaks, but the mammato and cumuliform growths are most pronounced when the clouds occur in the medium levels, where the temperature changes due to fallstreak growth and evaporation are greatest, and where the lapse rate sometimes substantially exceeds the saturated adiabatic. The fallstreak regeneration then leads to the production of altocumulus castellatus resembling miniature cumulonimbus, often arranged in groups lying along the direction of the wind shear. It is interesting that the release of instability in the altocumulus levels is usually effected by wave motions, superimposed on a large-scale vertical motion which gradually brings the air towards saturation. Eventually the wave motions cause the appearance of patches of altocumulus clouds, often of clearly lenticular form, which sprout cumuliform towers, and, if occurring at appropriate temperatures, produce fallstreaks which concentrate convective development in separate persistent groups or lines of clouds.

Clouds of this form are mentioned by Höhndorf (1947) to provide opportunities for sailplane soaring. He describes a winter occasion when inclined fallstreaks of granular snow and graupel extended between about 3,000 m and 1,500 m. 'Above the lower ends the air ascended in the neighbourhood of the fallstreaks, and reached upward speeds of up to 2 m/sec for a period up to almost an hour. A new cloud formed at the head of the updraught beside the old, sinking one, which after some time gave further ice precipitation, which produced a third cumulus, and so on.'

Conclusion

Flocciform or shred-clouds, sometimes becoming cumuliform, frequently appear over well-developed fallstreaks of ice crystals, evidently when the crystals fall through strongly ice-supersaturated air in which the release of latent heat causes warming of a few tenths of a degree Centigrade.

Convective bubbles of the warmed air rise out of the crystal trails and, particularly at the higher cirrus temperatures and when there is a steep lapse rate, form fresh clouds after an ascent of a few hundred metres. The effects of this process are dependent upon the temperature regime and the lapse rate.

At temperatures above about — 12°C fallstreaks are very rare. At temperatures between about — 12°C and — 30°C fallstreaks are common beneath isolated clouds predominantly composed of supercooled droplets. In normal lapse rates, close to the wet-adiabatic, the process can renew the cloud and thus regenerate the fallstreaks. In steeper lapse rates a very characteristic and persistent form of altocumulus castellatus is produced, with dense and mammilated fallstreaks.

At temperatures near — 40°C fallstreaks are still very common, being the typical form of isolated cirrus, but appear independently of any droplet mother-cloud, apparently forming on special, rare ice nuclei at near-saturation with respect to liquid water. In this temperature-regime the convective process can regenerate the fallstreaks without the formation of a special mother-cloud; shred-cloud formation is common, however, and iridescence in the shreds occasionally shows them to contain droplets. At the lower temperatures most of these droplets are likely to freeze, and crystals are thus produced in concentration of order $10/cm^3$, much in excess of those occurring in dense fallstreaks; consequently the shred-clouds do not always become transformed into fallstreaks. In very steep lapse rates convective cirrus develop which have forms analogous to the altocumulus castellatus previously mentioned.

At temperatures below about — 50°C the warming of air in fall-streak trails becomes very small; fallstreak regeneration therefore becomes unlikely, and, even in very steep lapse rates, shred-cloud formation still less likely. At very low temperatures the concentration of water condensed in cirrus becomes so small that the fallstreak is not the characteristic cloud form.

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Airflow over mountains: the lee-wave amplitude

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Summary

Although orographic lee waves are probably common over the British Isles, their amplitude is so critically dependent on airstream characteristics and on the scale of local topography that they are likely to be insignificant unless a variety of conditions is satisfied. Precise criteria for large-amplitude waves cannot be formulated for airstreams in general, but simplification of the general problem illustrates the pronounced effect of mountain size and shape, stability and wind speed on the lee-wave amplitude.

1. Introduction

In theoretical studies of airflow over mountains a good deal of attention has been given to formulating and discussing conditions under which a system of atmospheric lee waves will form in an airstream downwind of hills or mountain ranges (e.g., Scorer 1949 and 1953). These conditions involve the vertical variation of a parameter usually denoted by l^2 and given by

$$l^2 = \frac{g\beta}{U^2} - \frac{1}{U}\frac{\partial^2 U}{\partial z^2}$$

where g = acceleration due to gravity

β = stability factor defined as $\frac{1}{\theta}\frac{\partial\theta}{\partial z}$, θ being the potential temperature

U = component of wind across the mountain ridge

z = height

Theoretical study has shown that an airstream can contain lee waves only if l^2 decreases sufficiently with height. In practice such a simple criterion is often difficult to exploit owing to the complex, irregular vertical distributions of l^2 usually found in real airstreams. It does appear that lee-wave conditions are frequently satisfied over the British Isles, especially during winter, but familiarity with aircraft reports has given us the impression that only a few of these theoretically possible waves have substantial effects on powered aircraft. This suggests that, although waves may be very common, their amplitude may often be so small that pilots scarcely notice their effects during flight. On some occasions, however, pilots have encountered waves of distinctly troublesome magnitudes.

Accordingly, a preliminary study of the factors controlling the lee-wave amplitude has been made and this paper reveals the extreme sensitivity of such amplitude to the stability characteristics of the general airstream, and to the local scale of mountain ranges in relation to the wavelength of the free oscillations of the airstream.

2. Factors determining the lee-wave amplitude

Scorer (1949) has derived expressions for the flow over an idealized two-dimensional ridge whose height, ζ, at a horizontal distance x from the crest is given by

$$\zeta = \frac{hb^2}{b^2 + x^2} \quad . \quad . \quad . \quad . \quad (1)$$

where h is the height of the ridge and b is the ' half-width ' parameter. The flow pattern through lee waves induced by such a ridge is indicated by the term

$$\zeta_z = -2\pi h\, be^{-kb}\,(U_1/U_z)\,\psi_{z,k}\,(\partial\psi_{1,k}/\partial k)^{-1}\sin kx \quad . \quad . \quad (2)$$

where ζ_z = displacement of a streamline from its undisturbed level at height z

k = the lee-wave number

U = horizontal wind speed

ψ = satisfies the equation

$$\frac{\partial^2\psi}{\partial z^2} + (l^2 - k^2)\,\psi = 0 \quad . \quad . \quad . \quad . \quad (3)$$

and the suffix 1 refers to ground level.

An attempt to assess the relative importance of mountain size, temperature and wind profiles on wave amplitude could be made by examining each of the displacement factors in turn, but here we consider it convenient to study separately the three factors,

$$hbe^{-kb}, \quad U_1/U_z \quad \text{and} \quad \psi_{z,k}\,(\partial\psi_{1,k}/\partial k)^{-1}.$$

The effect of U_z could be eliminated from discussion by considering vertical speeds directly, but it is not difficult to assess such vertical speeds after visualizing the wave pattern, and the wave pattern itself is amenable to simple physical perception.

3. The effect of mountain size and shape

So far as the mountain is concerned, the lee-wave amplitude is proportional to hbe^{-kb}. A linear variation with mountain height is to be expected since the results are based on perturbation theory. The expression be^{-kb} has a maximum when $b = k^{-1}$; this means that if the mountain height remains constant the lee-wave amplitude attains its greatest value when the mountain width is suitably adjusted to the natural wavelength of the airstream. The variation of lee-wave amplitude with mountain width is illustrated by the dashed curve in Fig. 1. In this diagram k has been taken as 1 km^{-1}, i.e., the wavelength of the airstream is 2π km. The maximum at $b = k^{-1}$ is pronounced, and the amplitude falls off sharply for narrower or broader mountains. We may liken this effect to resonance; the natural wavelength is a function of the l^2 profile, i.e., of the airstream's stability and wind profiles, but the lee wave amplitude is likely to be small if the width of the mountain does not match the wavelength of the airstream.

We next enquire how the lee-wave amplitude varies if we vary the mountain size while preserving its shape, i.e. the height and width are varied in the same proportion. On this basis the variation of lee-wave amplitude has the same form as previously but the maximum amplitude now occurs when $b = 2k^{-1}$; the appropriate curve is drawn as a full line in Fig. 1. Thus we can regard the mountain for which $b = k^{-1}$ as having the optimum width for any specified height; if we then increase the overall mountain size the lee-wave amplitude tends to increase with the height factor, h, and to decrease as the width departs from the optimum value. Keeping the ratio $h : b$ constant allows the increasing height to dominate the amplitude tendency until $b = 2k^{-1}$; any further increase in overall mountain size produces a sharp decline in lee-wave amplitude. Indeed we see, for example, that if a given airstream produces lee waves of a certain amplitude downstream of a mountain for which $b = 2k^{-1}$ then the same airstream would only produce waves about 1/27 of that amplitude downstream of a mountain four times larger. Clearly this ' resonance ' effect is of considerable importance.

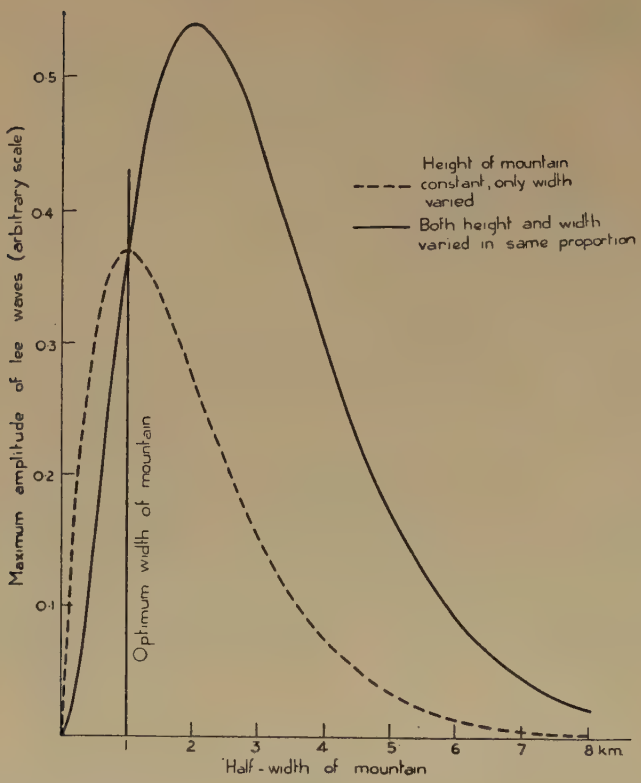

Figure 1. Variation of lee-wave amplitude with size of mountain ($k = 1$ km^{-1}).

We infer that any large-amplitude lee waves which happen to occur in an airstream will not necessarily be associated with the largest of the mountain ridges or hills in the underlying terrain. Thus an airstream which is favourable for significant waves over, say, the Cotswolds or the Chilterns will not necessarily produce even bigger waves over the Scottish Highlands or the Welsh Mountains.

We must, however, be cautious about exploiting generalizations deduced in this way because typical hilly terrain is irregular and contains orographic features on all scales ; we are not, in practice, dealing with mountains of geometrically similar shape (for example, a very large mountain ridge of long overall wavelength may possess a steep slope of, effectively, much shorter wavelength). There will, therefore, be many exceptions to the general rule for large mountains to require airstreams having a long wavelength for the maximum effects to be realized. Nevertheless, Förchtgott concluded, from his observations in Czechoslovakia (1949), that the larger the mountains the stronger are the wind speeds necessary for large-amplitude waves, and, of course, high wind speeds are theoretically associated with long wavelengths.

4. Effect of the wind profile

The lee-wave amplitude varies as U_1/U_z where U_1 is the surface wind speed and U_z the wind at height z. Consideration of this factor alone may give the impression that a decrease of wind with height would favour large-amplitude lee waves. But we know that for waves to be possible at all l^2 must decrease with height. Such a decrease can be achieved by an increase of wind speed or by a decrease of stability with height, or by some combination of wind and stability distributions. We conclude that if we have two

airstreams which, although having different wind and stability profiles, nevertheless have the same l^2 profiles, then the airstream in which U_1/U_z is larger will have the greater-amplitude lee waves. Another way of looking at this is to note that, although an airstream with very light winds near the surface and much stronger winds higher up can contain lee waves, the amplitude of the waves would probably be small ; other things being equal, larger-amplitude waves would occur if the airstream satisfies the conditions for waves partly or mainly by stability variations, so that moderate or strong surface winds can be present without the necessary variation of l^2 with height being destroyed.

This viewpoint is consistent with the evidence from aircraft reports collected at Northolt Airport and discussed by Pilsbury (1955). In these reported waves marked stability was always present at middle levels while the low-level wind speeds were usually more than 20 kt and never less than 15 kt.

5. Effect of the l^2 profile

The lee-wave amplitude also varies as $\psi_{z,k}\,(\partial\psi_{1,k}/\partial k)^{-1}$, which is a function of height and the parameter l^2. At one or more levels this amplitude factor attains some maximum value and we shall focus our attention on this maximum as a basis for studying the general effect of the l^2 profile. Variations in the l^2 profile will of course modify the lee-wave number k and also, therefore, the factor be^{-kb}. However, in the present section we confine ourselves to examining the factor $\psi_{z,k}\,(\partial\psi_{1,k}/\partial k)^{-1}$ alone. For a given mountain the simultaneous variations of be^{-kb} would intensify the characteristics which emerge. For ease of mathematical manipulation we consider the atmosphere to comprise :

(a) an adiabatic layer ($l = 0$) from the ground to a height h_a.

(b) an intermediate layer in which $l = l_i$ between the levels h_a and $(h_a + h_i)$.

(c) an upper layer in which $l = l_s$ ($< l_i$) extending upwards from the level $(h_a + h_i)$.

Although this three-layer structure is a compromise between reality and simplicity it is broadly in accordance with the type of airstream structure which obtains on occasions when aircraft report marked waves. In the first part of this discussion we carry the simplification further by putting $h_a = 0$. Then, taking the top of the intermediate layer as the zero of the height scale (i.e. the ground at $z = -h_i$) and assuming that waves must die out at high levels, the appropriate solutions of Eq. (3) are

$$\psi_s = A \exp(-\mu_s z) \qquad (4)$$

$$\psi_i = A\left\{\cos\mu_i z - \frac{\mu_s}{\mu_i}\sin\mu_i z\right\}$$

where $\mu = \{|l^2 - k^2|\}^{\frac{1}{2}}$, the suffixes i and s refer to the lower and upper layers respectively, and A is a constant.

If we apply the condition that ψ_1 must vanish at the ground downwind of an obstacle, we obtain an equation determining the lee wave number, k. The equation may have a number of solutions, but, as Scorer (1949) has pointed out, the smallest root (corresponding to the longest wavelength) will dominate the flow. If the wave equation has n solutions we find that, at the n heights above the ground given by $h_n = \pi/2\mu_i$, $3\pi/2\mu_i$ . . . . $(2n-1)\,\pi/2\mu_i$, the amplitude factor $\psi_{z,k}\,(\partial\psi_{1,k}/\partial k)^{-1}$ assumes a maximum value, a_n, given by

$$a_n = \frac{\mu_i}{k}\left(h_i + \frac{1}{\mu_s}\right)^{-1} \qquad (5)$$

Mere inspection of this equation does not easily reveal how a_n varies with the airstream parameters l_i, l_s and h_i; this is because it is not possible to express k explicitly

in terms of these parameters. It is simpler to study graphical representations of the equation and for this purpose the introduction of an additional dependent variable is helpful. Thus as k must lie between l_i and l_s we may write

$$k^2 = l_s^2 + L^2 \cos^2 \theta \quad . \quad . \quad . \quad . \quad . \quad (6)$$

$$L^2 = l_i^2 - l_s^2 \quad . \quad . \quad . \quad . \quad . \quad (7)$$

where $0 \leqslant \theta \leqslant \pi/2$

We shall then have

$$\left.\begin{aligned} \mu_i &= L \sin \theta \\ \mu_s &= L \cos \theta \end{aligned}\right\} \quad . \quad . \quad . \quad . \quad . \quad (8)$$

and the equation for the lee wavelength reduces to

$$(n\pi - \theta)/\sin \theta = Lh_i \quad . \quad . \quad . \quad . \quad (9)$$

With this notation the lee-wave amplitude factor at the levels where it is a maximum becomes

$$a_n = L^2 (l_s^2 + L^2 \cos^2 \theta)^{-\frac{1}{2}} (n\pi - \theta + \tan \theta)^{-1} \sin^2 \theta \quad . \quad . \quad (10)$$

This equation may be used to calculate the maximum amplitude for various values of l_s, L and θ, the corresponding values of h_i being determined by Eq. (9). Fig. 2 shows

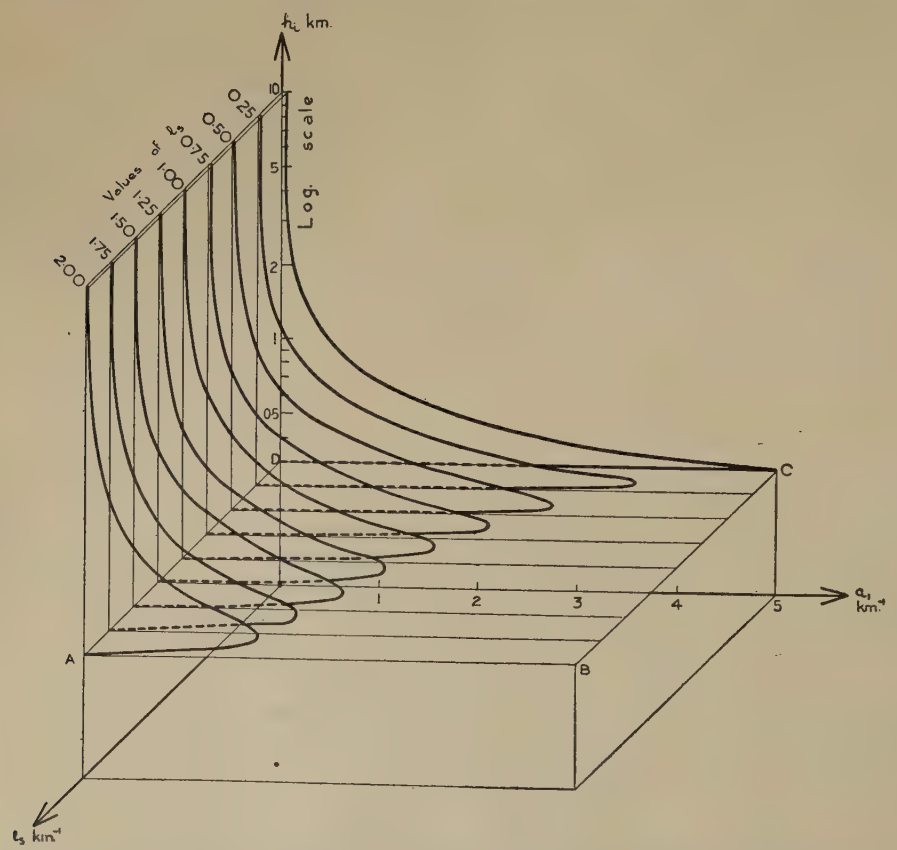

Figure 2. Variation of lee-wave amplitude (a_1) with airstream characteristics (L = constant = 5 km^{-1}).

an isometric sketch in which the family of curves delineates a surface illustrating the way in which the maximum lee-wave amplitude varies with l_s and h_i for a constant value of L ($= 5\ \text{km}^{-1}$ in this figure). For values of h_i below the plane A B C D the condition for lee waves is not satisfied. A striking feature of the diagram is that as h_i is increased beyond this critical value we pass very rapidly from a condition in which there can be no lee waves to one in which they will have large amplitudes. It can also be demonstrated that if h_i were kept constant as L varied, the lee-wave amplitude would behave similarly as a limiting value of L was crossed. It has been stated that the greater the decrease of l^2 with height, the more likely are lee waves because one can then be more confident that the condition for them is satisfied. Although this statement is correct it is liable to imply that the greater the decrease in l^2, the larger will be the amplitude of lee waves; casual interpretation of the perturbation restrictions may also suggest that deep layers having high values of l favour large-amplitude waves. However, it appears that the optimum conditions, when the airstream is most sensitive, are very close to the critical conditions for waves, and that if conditions are made ' more favourable ' for waves to occur the amplitude falls off sharply. We also note that as l_s increases, this sensitivity decreases somewhat (but remains important) and the amplitude falls off.

The variations illustrated in Fig. 2 are for a constant value of L of $5\ \text{km}^{-1}$. We next examine the effect of varying L, which is a measure of the decrease of l^2 with height. Fig. 3 shows curves for the lee-wave amplitude (once again for the level at which it is a maximum) for $L = 2$, 5 and $10\ \text{km}^{-1}$. For these curves the other variable is h_i; l_s is kept constant at $0{\cdot}2\ \text{km}^{-1}$ throughout. The corresponding wavelengths are depicted in the left-hand section of the diagram. Full curves refer to $n = 1$, and dashed curves to $n = 2$; curves for higher values of n are not shown, but the tendency for the wave amplitudes to decrease as k acquires more possible values is maintained, and unless l_s is extremely small this decrease is a rapid one.

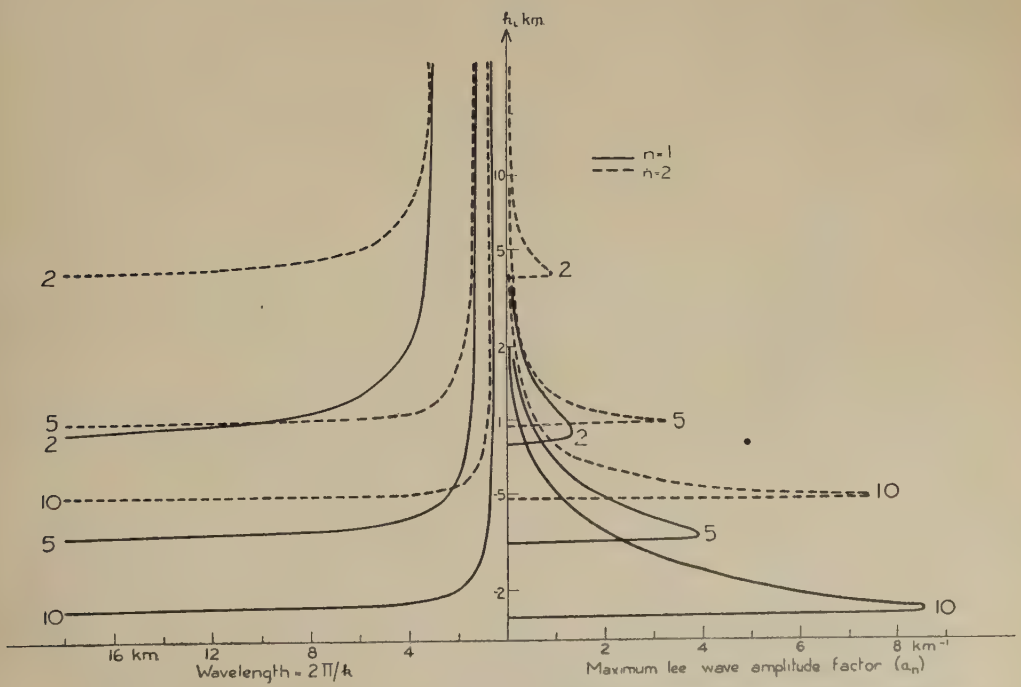

Figure 3. Variation of lee-wave amplitude and wavelength with airstream characteristics (l_s = constant = $0{\cdot}2\ \text{km}^{-1}$; values of L (2, 5 and $10\ \text{km}^{-1}$) are marked against curves).

We see again the great sensitivity of the airstream near the critical values of the parameters for waves to occur; it is also apparent that the maximum amplitude becomes potentially greater as L is made larger. Indeed, it appears that the more the lower layer tends towards a sharp inversion, the greater does the maximum possible amplitude become. However, since the boundary conditions used at the interface between the two layers assume continuity of density we cannot extend our conclusions to an ideally sharp inversion without modifying these boundary conditions. Nevertheless the conclusion remains valid that greater amplitudes are possible if there is considerable static stability through a shallow layer than with smaller stability through a deep layer.

The variation of amplitude with L illustrated in Fig. 3 enables us to visualize a set of surfaces, one of which is given in Fig. 2, for a single value of L. We may then summarize the principal conclusions of Figs. 2 and 3 by saying that the largest-amplitude lee waves will be possible when

(a) conditions are slightly more favourable for waves than the limiting conditions for them to occur at all. By ' slightly more favourable ' we mean that the stability or depth of the layer having the large l^2 should be a little greater than the limiting combination for waves.

(b) L should be large.

(c) l_s should be small.

We next consider briefly the maximum vertical velocity in these lee waves; its variation with the airstream parameters will not follow exactly that of the amplitude because of the wavelength changes. Well away from the mountain the displacement of the streamline at the level where $\psi_{z,k}\,(\partial\psi_{1,k}/\partial k)^{-1}$ has a maximum will be given by putting $z = h_n$ in the equation

$$\zeta_z = -\,2\pi h\, be^{-kb}\,(U_1/U_z)\,a_n \sin kx \qquad . \qquad . \qquad . \qquad (11)$$

so that if w is the magnitude of the maximum vertical velocity,

$$w = [U_z\,(\partial\zeta_z/\partial x)]_{\max} = 2\pi h\, be^{-kb}\,U_1\,a_n\,k \qquad . \qquad . \qquad (12)$$

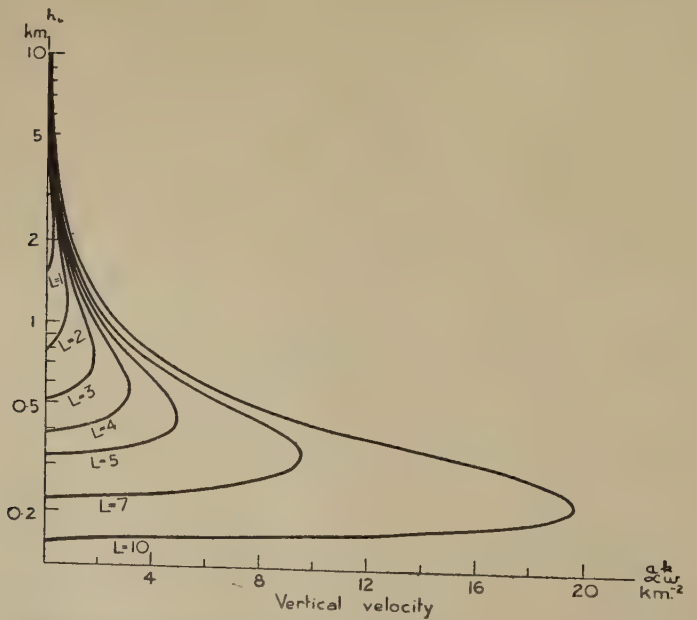

Figure 4. Variation of maximum vertical velocity in lee waves with airstream characteristics. Values of L are indicated. Results are independent of l_s.

Thus
$$w \propto \mu_i \left(h_i + \frac{1}{\mu_s}\right)^{-1}$$
$$\propto L^2 (n\pi - \theta + \tan\theta)^{-1} \sin^2\theta \quad . \quad . \quad . \quad (13)$$

We note that Eq. (13) is independent of l_s so that a family of curves is sufficient to illustrate the variation of vertical velocity with the three parameters defining the airstream. Such curves are drawn in Fig. 4 for $L = 1$, 2, 3, 4, 5, 7 and 10 km^{-1} and for $n = 1$. These curves exhibit features which are similar to the amplitude curves, the main differences being that sensitivity is somewhat less critical near the optimum combination of h_i and L.

It is impracticable to extend the graphical representation of the maximum amplitude factor to deal with a three-layer atmosphere fully, but we can study such a model to a limited extent. The addition of an adiabatically mixed layer, that is, a layer with neutral static stability, (denoted by suffix a) at the bottom requires an additional solution of Eq. (3) having the form
$$\psi_a = B \cosh kz + C \sinh kz \quad . \quad . \quad . \quad (14)$$

where B and C are constants. Further application of the boundary conditions leads to the following expression for the lee-wave amplitude factor at the level where it is a maximum

$$a_n = (\mu_i^2 - l_i^2 \sin^2\gamma)^{\frac{1}{2}} \left[k\left(h_i + \frac{1}{\mu_s}\right) + \tfrac{1}{2}(l_i^2/Lk)\sin 2\gamma \operatorname{cosec}\theta - (\mu_i^2 - l_i^2\sin^2\gamma)\, h_a/k\right]^{-1} \quad (15)$$

where
$$\tan\gamma = (\mu_i/k)\tanh(kh_a) \quad . \quad . \quad . \quad . \quad (16)$$

and
$$\gamma = n\pi - \theta - L\, h_i \sin\theta \quad . \quad . \quad . \quad . \quad (17)$$

The condition for these equations to yield n real solutions for k can be derived as
$$L\, h_i > \tfrac{1}{2}(2n - 1)\,\pi - \alpha \quad . \quad . \quad . \quad . \quad (18)$$

where
$$\alpha = \tan^{-1}[(L/l_s)\tanh(l_s\, h_a)] \quad \text{and} \quad 0 \geqslant \alpha \geqslant \pi/2 \quad . \quad . \quad (19)$$

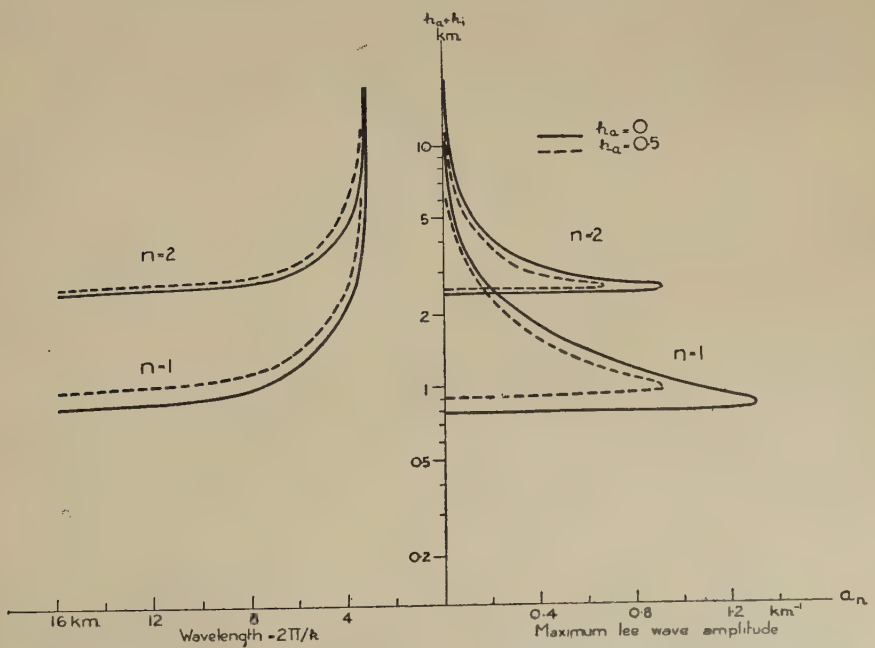

Figure 5. Variation of lee-wave amplitude and wavelength for three-layer airstream ($L = 2$ km^{-1}; $l_s = 0{\cdot}2$ km^{-1}).

Eq. (18) may be regarded as the condition for lee waves to be possible in this three-layer model and it corresponds to $L\,h_i > \frac{1}{2}(2n-1)\,\pi$ for the two-layer model; indeed as $h_a \to 0$ the former condition tends to the latter. We note that it is possible for the addition of an adiabatically mixed surface layer to make lee waves possible when they would not occur otherwise, provided the adiabatic layer is not added at the expense of the intermediate layer. The effect on the maximum amplitude of the additional layer can be seen in Fig. 5 which contains amplitude and wavelength curves for $L = 2\ \text{km}^{-1}$ and $l_s = 0{\cdot}2\ \text{km}^{-1}$. This figure contains dashed curves computed for an airstream with a low-level adiabatic layer of depth 0·5 km and full curves applicable when the adiabatic layer is absent. The general conclusions reached for the two-layer model are seen to be equally applicable if there is an additional layer at the surface having an adiabatic lapse rate, the only differences being a limited reduction in amplitude and an increase of wavelength throughout the range of values.

6. Conclusion

Over the British Isles airstreams which formally satisfy the theoretical conditions for lee waves are probably very common, especially in winter. However, the amplitude of such waves will be subject to very wide variations even amongst airstreams having very similar profiles of l^2 because of the critical way in which the amplitude depends on the airstream characteristics. The largest-amplitude waves occur when the airstream satisfies the condition for waves by only a small margin, and in this region very large changes of amplitude may result from very small changes in the airstream characteristics. Apart from the question of this sensitive region, it may be said that larger-amplitude waves are theoretically more likely in airstreams containing a shallow layer of great stability than in conditions of slight stability through a deep layer. The main effect of an adiabatically mixed layer near the ground is to reduce the amplitude and increase the wavelength.

Quite apart from the variations which depend on the airstream characteristics it appears that a given airstream is sensitive to the scale of the disturbing mountain. If the mountain is made broader or narrower than the optimum width for the prevailing airstream a considerable reduction of wave amplitude is to be expected.

Acknowledgment

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The representation of the annual temperature variation over central and northern Europe by a two-term harmonic form

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SUMMARY

A harmonic analysis is made of the annual temperature variation at 42 stations spread over central and northern Europe, using in each case 5-day mean temperatures averaged over a considerable number of years. It is found that the first two harmonics account for nearly all the annual variance, the residual variance being usually less than 0·4 per cent of the total. Most of this residual variance seems to be random in origin, but a small part may be due to systematic departures from the two-term harmonic form.

1. INTRODUCTION

This work is an example of the process, familiar in all the physical sciences, of fitting a multitude of observations into the framework of a concise and easily understood mathematical law.

The actual process, that of representing the annual temperature variation by some of the terms of a Fourier series, seems to have been used first by Bessel (1828), and since then has been used by many authors, for example Forbes (1860), Plantamour (1867), Brooks and Mirrlees (1930), Pollak (1947) and Reynolds (1955). Fourier analysis may be looked on either as a mathematical process which if carried far enough will produce an exact representation of the raw data, or as a physical process designed to filter the significant information from the data while discarding the irrelevant 'noise.' Work having the second object will use as few harmonics as possible, consistent with the residuals having the haphazard and non-recurrent character associated with 'noise.' This paper is concerned with the goodness of fit obtainable by a two-term harmonic form (having five independent parameters) and with the question whether the residuals left after removing from each record the best fitting two-term harmonic form show any regularity or systematic variation with year, place or season.

2. METHOD

We propose to fit the annual temperature variation by an expression of form :

$$T = a_0 + \sum_{j=1}^{N} (a_j \cos j\,\theta + b_j \sin j\,\theta) \quad . \quad . \quad . \quad (1)$$

where θ increases by 2π during the year. θ can be interpreted astronomically as the sun's geocentric longitude, but as a matter of practical convenience we measure θ from zero at midnight on 31 December and treat all years as having exactly 365 days, ignoring 29 February in leap years, Indeed, the form of the available raw data gives us no option.

If we have a series of observations covering a year with variance $V(T)$, then by taking N large enough in Eq. (1) we can fit the series as closely as we please, and by a standard theorem

$$V(T) = \tfrac{1}{2}(a_1^2 + b_1^2 + a_2^2 + b_2^2 + \ldots + a_N^2 + b_N^2) \quad . \quad . \quad (2)$$

Thus, for example, the essentially positive expression $V(T) - \frac{1}{2}(a_1^2 + b_1^2)$ measures the sum of half the squared amplitudes of all the harmonics except the first, and hence

the mean-square discrepancy between the curve representing the observed values of T and that representing the first harmonic. To test the goodness of fit of any desired combination of harmonics we need only compare the residual variance estimated in this way with the total variance $V(T)$. With the data used here it appears that when the first two harmonics are included the residual variance is too small to have much practical importance.

TABLE 1. THE STATIONS USED

Station	Lat. (N.)	Long. (E.)	Altitude (m)	Period over which data are averaged	Source of data
Augsburg	48°22′	10°54′	502	1851-1930	*Deutsche Klimakunde*, Bd. II, Tables, Rwd., Berlin, 1939
Bayreuth	49°57′	11°34′	364	1851-1930	
Berlin	52°31′	13°22′	62	1851-1930	
Breslau	51°07′	17°02′	147	1851-1930	
Brussels	50°52′	04°23′	39	1831-1860	Quetelet (1867)
Brocken	51°48′	10°37′	1,150	1896-1930	*Deutsche Klimakunde*
Darmstadt	49°33′	08°40′	150	1851-1930	
Dresden	51°04′	13°46′	112	1851-1930	
Edinburgh	55°57′	03°15′W.	76	1795-1804, 1821-1850	Forbes (1860)*
Emden	53°22′	07°12′	8	1851-1930	*Deutsche Klimakunde*
Erfurt	50°58′	11°04′	218	1851-1930	*Deutsche Klimakunde*
Frankfurt a. O.	52°20′	14°35′	57	1851-1930	
Friedrichshafen	49°39′	09°29′	408	1851-1930	
Glasgow	55°52′	04°17′W.	56	1868-1912	*Met. Off. Geophys. Mem.*, No. 23
Görlitz	51°10′	15°00′	217	1851-1930	*Deutsche Klimakunde*
Gütersloh	51°55′	08°23′	81	1851-1930	*Deutsche Klimakunde*
Halle	51°29′	11°59′	94	1851-1930	
Halle	51°29′	11°59′	94	1851-1900	Hoelscher (1954)
Halle	51°29′	11°59′	94	1901-1950	
Hohenpeissenberg	47°48′	11°01′	994	1851-1874, 1879-1930	*Deutsche Klimakunde*
Kew	51°28′	00°19′W.	8	1871-1950	Meteorological Office tabulation 1951
Kiel	54°19′	10°08′	47	1851-1930	*Deutsche Klimakunde*
Kleve	51°48′	06°08′	48	1851-1930	
Köln	50°56′	06°57′	56	1851-1930	
Königsberg	54°43′	20°30′	7	1851-1930	
Konitz	53°42′	17°34′	175	1851-1930	*Deutsche Klimakunde*
Köslin	54°12′	16°11′	46	1851-1930	
Lüneberg	53°15′	10°24′	20	1851-1930	
Memel	55°43′	21°08′	18	1851-1930	
München	48°10′	11°30′	531	1851-1930	
Posen	52°25′	16°56′	66	1851-1930	*Deutsche Klimakunde*
Potsdam	52°23′	13°04′	80	1901-1950	Branicki (1952)
Prague	50°05′	14°25′	197	1775-1839	Hlaváč (1937) worked in Meteorological Office
Prague	50°05′	14°25′	197	1840-1934	Hlaváč (1937)
Ratibor	50°06′	18°13′	191	1851-1930	*Deutsche Klimakunde*
Schneekoppe	50°44′	15°44′	1,618	1881-1930	*Deutsche Klimakunde*
Stettin	53°26′	14°30′	26	1851-1930	
Stuttgart	48°47′	09°10′	267	1851-1930	
Tilsit	55°04′	21°54′	18	1851-1930	
Torgau	51°34′	13°00′	99	1851-1930	
Treuberg	54°02′	22°30′	162	1851-1930	*Deutsche Klimakunde*
Trier	49°45′	06°38′	150	1851-1930	
Uppsala	59°51′	17°38′	24	1871-1950	*Bull. Mens. Observ. Met. Uppsala*, **1-82**, worked in Meteorological Office
Vienna	48°15′	16°22′	203	1873-1947	*Zentr. Met. Geodynam. Jahrb.*, 1783-1947, Vienna, worked in Meteorological Office
Zugspitze	47°25′	10°59′	2,962	1901-1930	*Deutsche Klimakunde*

* Observations made at three locations less than 1¾ mi apart. Elevation from 70-59 m.

3. Data

For completeness, the investigation should start with daily mean temperatures for every day of the year. However, an investigation which started on these lines would be handicapped by the labour involved, and by the difficulty of finding enough suitable data. There are, however, many publications containing mean temperatures for the 73 non-overlapping 5-day periods which cover the year, and it is unlikely that any temperature anomaly which is smoothed out by averaging over a period as short as five days will recur on the same day from year to year. So the present investigation was based on 5-day mean temperatures in the expectation that any significant and annually recurring feature of the sequence of daily mean temperatures would also be present in the 5-day means. Table 1 gives particulars of the stations used, and the sources from which the data were obtained.

4. Calculation

The calculation of harmonic coefficients from 73 equally-spaced observations is not difficult, but it is laborious and hard to check. The formulae are

$$a_k = \tfrac{2}{73} \sum_{j=1}^{73} T_j \cos k\,\theta_j \quad . \quad . \quad . \quad . \quad (3)$$

and

$$b_k = \tfrac{2}{73} \sum_{j=1}^{73} T_j \sin k\,\theta_j \quad . \quad . \quad . \quad . \quad (4)$$

where $k = 1$ or 2 and T_j, θ_j are the observed mean temperature and the angle corresponding to the j-th 5-day period of the year. Here θ_1 is the angle corresponding to the middle of the first 5-day period, 1-5 January. The values of $\cos \theta_j$, $\sin \theta_j$ are given by Pollak (1947) and these were multiplied by the factor $\frac{2}{73}$ and incorporated in a special computing form prepared for the purpose. The 5-day mean temperatures are written into another form which also contains spaces for all numbers required in calculation. Corresponding columns on the two forms are brought together and the products are formed and summed on a desk machine. Experience shows that arranged in this way the calculation is reasonably easy to check. For about 8 stations the harmonic analysis was carried out in this way; later, data for 34 more stations were analysed by Hollerith methods (on a Junior Rolling Tabulator) taking advantage of facilities for processing the cards offered by Rothamsted Experimental Station. In Table 2 are given the coefficients of the first two harmonics for each set of data, the total annual variance, and the residual variance after subtracting that due to the first two harmonics.

5. The geographical variation of the harmonic parameters

The semi-amplitude and phase of the first and second harmonics are shown in Figs. 1 to 4. Besides the parameters calculated in the present work values have been obtained from two other sources. The parameters for 18 Norwegian stations are given by Birkeland (1936) while those for 160 stations in the British Isles have been calculated by the Meteorological Office. These values which are computed from monthly mean temperatures are somewhat less reliable than values computed from 5-day mean temperatures. Fuller information about the stations used by Birkeland is given in Tables 1 (a) and 2 (a).

Fig. 1 shows the semi-amplitude of the first harmonic. The amplitude increases generally towards the east, but there is a striking difference between the rapid increase as the Norwegian mountains are crossed and the much slower increase across the North German plain. In Fig. 2 the phase of the first harmonic is given by the calendar date on which the maximum is attained.

The German data show with surprising consistency that the phase is earlier in the interior of the Continent than in coastal regions. This association is borne out by the dense network of stations over the British Isles, which shows, moreover, that the change

TABLE 2. HARMONIC COEFFIEICENTS; TOTAL AND RESIDUAL VARIANCES

Station	Annual mean temperature and harmonic coefficients in degrees Celsius					Annual variance	Residual variance	Residual variance as percentage of total	No. of years
	a_0	a_1	b_1	a_2	b_2				
Augsburg	8·12	− 9·54	− 2·59	0·33	0·45	49·21	0·16	0·33	80
Bayreuth	7·76	− 9·42	− 2·55	− 0·22	0·28	47·81	0·12	0·25	80
Berlin	9·27	− 9·35	− 2·73	0·11	0·28	47·59	0·14	0·29	80
Breslau	8 58	− 9·88	− 2·94	− 0·06	0·16	53·25	0·13	0·24	80
Brussels	10·28	− 7·87	− 2·66	− 0·02	0·33	34·80	0·23	0·66	30
Brocken*	2·63	− 6·83	− 2·92	0·41	0·26	27·86	0·16	0·57	35
Darmstadt	9·61	− 8·76	− 2·22	− 0·05	0·36	40·93	0·06	0·15	80
Dresden	9·14	− 9·04	− 2·71	0·00	0·29	44·71	0·18	0·40	80
Edinburgh	8·31	− 5·68	− 2·14	0·21	0·48	18·65	0·08	0·43	40
Emden	8·52	− 7·65	− 2·85	0·15	0·43	33·51	0·12	0·36	80
Erfurt	7·93	− 8·94	− 2·61	− 0·07	0·24	43·58	0·19	0·44	80
Frankfurt a. O.	8·54	− 9·50	− 2·77	0·07	0·24	49·19	0·17	0·35	80
Friedrichshafen	8·84	− 9·21	− 2·65	− 0·13	0·26	46·08	0·12	0·26	80
Glasgow	8·29	− 5·10	− 2·04	0·45	0·54	15·36	0·05	0·33	45
Görlitz	8·24	− 9·43	− 2·83	− 0·41	0·21	48·70	0·18	0·37	80
Gütersloh	9·05	− 8·11	− 2·49	0·16	0·29	36·22	0·17	0·47	80
Halle (1851-1930)	9·07	− 9·22	− 2·63	0·07	0·30	46·12	0·15	0·33	80
Hohenpeissenberg*	6·00	− 8·11	− 2·91	0·12	0·43	37·49	0·23	0·61	76
Kew (1901-50)	10·29	− 6·23	− 2·39	0·34	0·62	22·63	0·10	0·44	50
Kiel	8·47	− 7·72	− 3·04	0·43	0·36	34·69	0·10	0·29	80
Kleve	9·13	− 7·75	− 2·45	0·11	0·41	33·26	0·14	0·42	80
Köln	10·24	− 8·00	− 2·44	− 0·02	0·38	35·19	0·15	0·43	80
Königsberg	7·06	− 9·90	− 3·47	0·22	0·16	55·20	0·15	0·27	80
Konitz	6·81	− 9·85	− 3·11	0·27	0·15	53·53	0·16	0·30	80
Köslin	7·22	− 8·82	− 3·21	0·26	0·30	44·21	0·14	0·32	80
Lüneberg	8·47	− 8·51	− 2·61	0·31	0·27	39·89	0·17	0·43	80
Memel	6·85	− 9·51	− 3·89	0·31	0·14	52·97	0·07	0·13	80
München	7·64	− 9·53	− 2·65	− 0·36	0·36	49·18	0·12	0·24	80
Posen	8·29	− 9·95	− 2·89	0·17	0·12	53·86	0·16	0·30	80
Potsdam	8·54	− 9·35	− 2·65	− 0·08	0·12	47·44	0·15	0·32	50
Prague (1775-1839)	9·57	− 10·33	− 3·01	− 0·35	0·08	58·16	0·18	0·31	65
Prague (1840-1934)	9·04	− 10·00	− 2·83	− 0·13	0·20	54·12	0·09	0·17	95
Ratibor	8·05	− 9·97	− 2·80	− 0·22	− 0·01	53·69	0·09	0·17	80
Schneekoppe*	0·24	− 7·28	− 3·07	0·36	0·19	31·60	0·28	0 89	50
Stettin	8·41	− 9·37	− 2·95	0·19	0·29	48·48	0·14	0·29	80
Stuttgart	10·00	− 8·98	− 2·35	− 0·16	0·41	43·34	0·13	0·30	80
Tilsit	6·59	− 10·53	− 3·40	0·32	− 0·05	61·41	0·14	0·23	80
Torgau	9·01	− 9·35	− 2·72	0·08	0·29	47·62	0·18	0·38	80
Treuberg	5·43	− 10·83	− 3·23	0·30	− 0·37	64·22	0·21	0·33	80
Trier	9·17	− 8·22	− 2·27	− 0·06	0·40	36·58	0·12	0·33	80
Uppsala	5·25	− 9·96	− 3·63	0·97	0·25	56·73	0·05	0·09	80
Vienna	9·38	− 10·09	− 2·77	− 0·50	0·19	54·89	0·06	0·11	75
Zugspitze*	− 4·92	− 6·10	− 3·19	0·22	0·16	24·12	0·39	1·62	30

* Mountain stations.

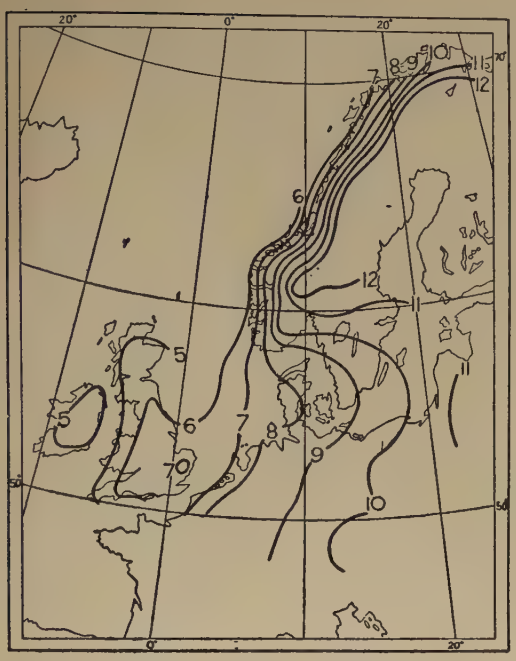

Figure 1. The semi-amplitude of the first harmonic (°C).

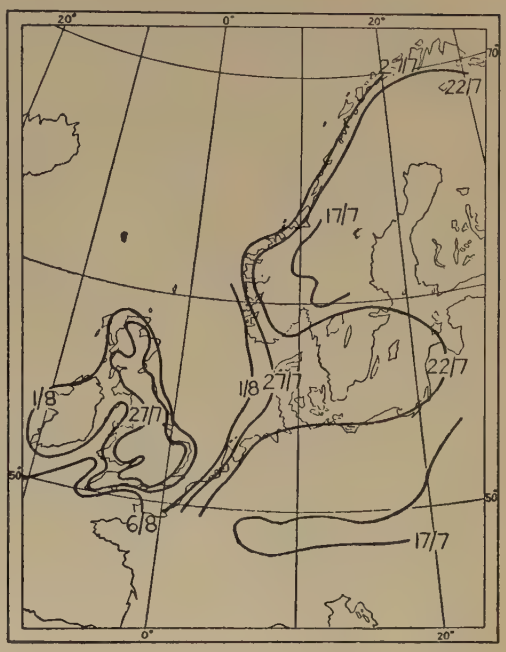

Figure 2. The phase of the first harmonic, date of maximum.

Figure 3. The semi-amplitude of the second harmonic (°C).

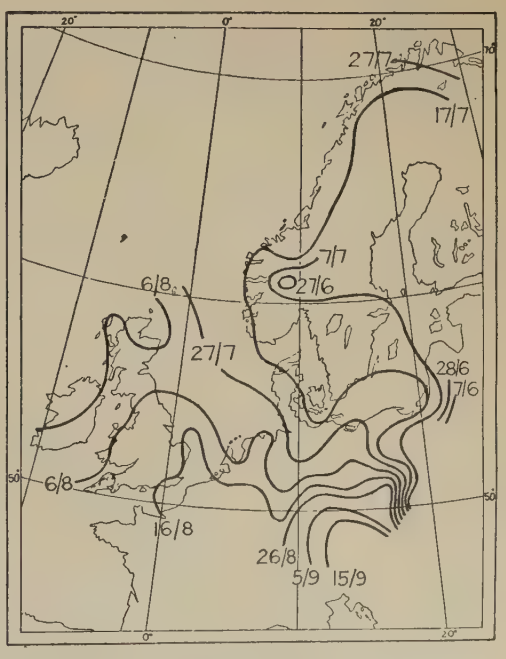

Figure 4. The phase of the second harmonic, date of maximum.

of phase is most rapid in a belt of perhaps 20 or 30 mi inland from the coasts. At the four mountain German stations the amplitude is smaller and the phase later than at nearby stations at lower altitudes. In these respects the mountain stations seem to have a more maritime climate than their neighbours. Figs. 3 and 4 give the semi-amplitude and phase of the second harmonic. The data are self-consistent enough to suggest that the variations shown are not due to chance, but we can hardly suggest from these data alone what the cause or large scale character of the variation may be.

TABLE 1 (a). THE NORWEGIAN STATIONS USED (after Birkeland, *Geofys. Publ.*, XIV, No. 1, 1936)

Station	Lat. (N.)	Long. (E.)	Altitude (m)
As	59°40′	10°46′	95
Bodo I.	67°17′	14°24′	22
Dalen I.	59°27′	07°58′	103
Dombas	62°05′	09°07′	643
Hamar	60°48′	11°04′	139
Hattfjelldal	65°36′	14°00′	235
Kristiansand N.	63°07′	07°45′	18
Kristiansand S.	58°10′	07°59′	22
Nordfjordeid	61°55′	06°00′	5
Oslo	59°55′	10°43′	22
Skudenes	59°09′	05°16′	2
Sorvaranger	69°44′	30°03′	5
Tromso I.	69°39′	18°58′	45
Trondheim	63°26′	10°25′	58
Ullensvang	60°19′	06°40′	10
Vollen	61°06′	08°58′	403

TABLE 2 (a). HARMONIC COEFFICIENTS, TOTAL AND RESIDUAL VARIANCES FOR NORWEGIAN STATIONS (after Birkeland, 1936)

Station	Annual mean temperature and harmonic coefficients in degrees Celsius					Annual variance	Residual variance*	Residual variance as percentage of total*	No. of years
	a_0	a_1	b_1	a_2	b_2				
As	4·96	− 10·10	− 3·24	0·65	0·18	56·44	0·04	0·07	60
Bodo I.	4·11	− 6·85	− 3·24	0·86	0·78	24·49	0·09	0·31	60
Dalen I.	5·07	− 9·90	− 2·98	0·75	0·32	53·75	0·06	0·11	40
Dombas	1·04	− 10·23	− 3·06	0·56	0·50	57·28	0·07	0·12	60
Hamar	3·44	− 11·19	− 3·56	0·48	− 0·03	69·07	0·04	0·06	52
Hattfjelldal	0·91	− 10·84	− 3·63	0·53	0·62	65·83	0·17	0·26	52
Kristiansand N.	6·59	− 5·39	− 2·74	0·60	0·58	18·63	0·03	0·16	60
Kristiansand S.	6·88	− 8·11	− 3·12	0·83	0·25	38·09	0·04	0·11	44
Nordfjordeid	6·03	− 7·25	− 2·27	1·14	0·48	29·82	0·17	0·57	15
Oslo	5·84	− 10·58	− 3·18	0·68	0·17	61·21	0·02	0·03	60
Skudenes	7·51	− 5·55	− 3·21	0·45	0·40	20·77	0·04	0·19	60
Sorvaranger	− 0·71	− 10·92	− 4·44	0·36	0·68	70·01	0·21	0·30	60
Tromso I.	2·40	− 6·79	− 3·15	1·09	0·90	29·02	0·05	0·17	60
Trondheim	4·78	− 8·14	− 2·86	0·63	0·58	37·57	0·06	0·16	51
Ullensvang	6·68	− 7·50	− 2·56	0·88	0·30	31·83	0·04	0·13	60
Vollen	1·89	− 11·44	− 3·56	0·08	− 0·01	71·81	0·17	0·24	60

* This value for the residual variance is half the sum of the squares of Birkeland's values for the semi-amplitudes of harmonics 3, 4, 5, and 6, and therefore is less than the true value.

6. The magnitude of the residual variance

The residual variance is given in Table 2 in absolute units ($°C^2$) and also expressed as a percentage of the total annual variance. It seems to be disproportionately large at three of the four mountain stations, and if we exclude these its value is never greater than 0·23 $°C^2$, and in some places is as low as 0·05 $°C^2$. Expressed as a percentage, the residual variance, except at the mountain stations, is generally less than 0·4 per cent. These figures give an upper limit to the order of magnitude of the errors we accept if we assume that the annual march of temperature is given by the best fitting two-term harmonic form.* For some purposes these errors may be negligible; for example, in long-range forecasting, the errors in predicting the future value of the temperature anomaly on a given day will probably far exceed the errors in predicting the normal temperature for that day. At any rate the investigator, knowing their probable magnitude, can decide whether to ignore them or not.

These results complete the main purpose of the paper, which is to give some idea what to expect if the annual temperature variation is assumed to follow the two-term harmonic form. The rest of the paper sets out to find a little more about the way in which the residual variance is made up. The data are not very suitable for this purpose, because the 5-day mean temperatures used have mostly been averaged at source, so that they give no information on variances between one year and another. Nevertheless they provide various indications which may serve as a basis for more elaborate investigations.

7. Departures from the two-term harmonic form at particular times of year

Even if departures from the two-term harmonic form are not important on the average in estimating the normal temperature they may be important at particular times of year, and they are of interest in themselves, even if very small, provided they can be shown to be systematic and permanent, since they may focus attention on periods of the year when synoptic situations peculiar to the season are to be expected.

The nature of these departures is shown in Fig. 5 which shows superimposed curves of the observed mean temperature and the best fitting two-term harmonic form for eight stations spread over the area covered by the observations. The horizontal scale which is the same for all stations gives the time of the year, while the vertical scales which are in degrees C have been separated to avoid confusion. Thus anomalies for different stations for the same period of the year appear in the same vertical line. It appears that the anomalies for any station for consecutive 5-day periods are serially correlated : in cases worked out later (Table 6) the first serial correlations are mostly between 0·5 and 0·7. There is also a suggestion of spatial correlation ; that is, the anomaly at a particular period of the year tends to have the same sign over a large part or the whole of the area for which we have data. These temporal and spatial correlations are what we would expect if the anomalies are due, not to fortuitous or local effects, but to large-scale features of the atmospheric circulation, and they must be allowed for when we decide on the statistical significance of the anomalies. For example, a single exceptional winter may produce a substantial negative contribution to the temperature anomalies for stations over a large area for several consecutive periods, and if we treat all these anomalies as independent we shall arrive at a wholly unjustified degree of confidence that similar anomalies will repeat themselves in future. However, to start with, we will examine the eight records to see what they suggest, without at this stage trying to show whether these suggestions are significant. The features which deserve further investigation are given in Table 3.

* The corresponding values in Table 2 (a) give for Birkeland's Norwegian data half the sum of the squares of the amplitudes of the third to sixth harmonics – they are mostly lower than our estimates of the total residual variance, as is to be expected.

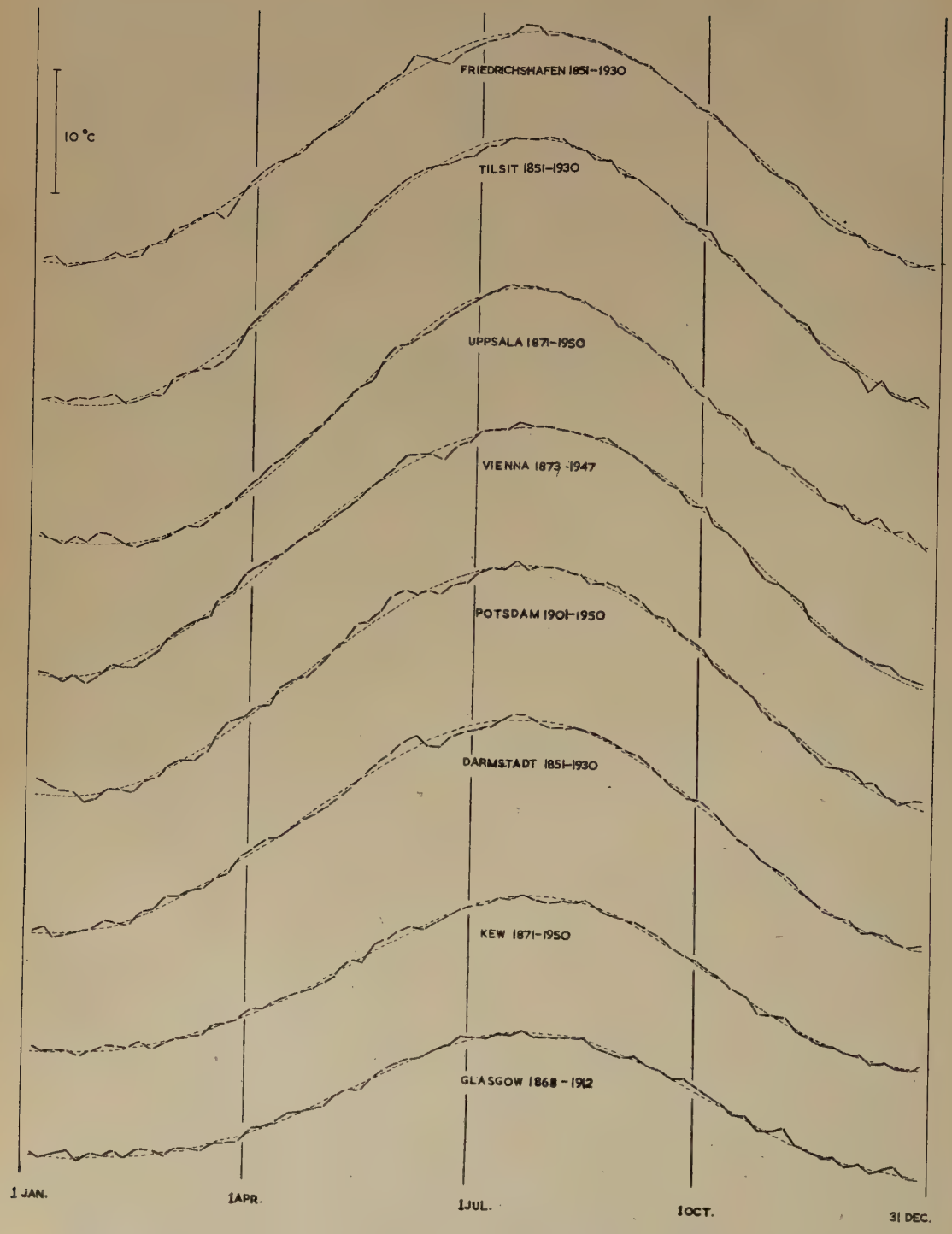

Figure 5. Anomalies of 5-day mean temperatures for 8 stations in north-west and central Europe.

TABLE 3. 5-DAY PERIODS IN WHICH THE TEMPERATURE ANOMALIES ARE REMARKABLY CONSISTENT IN SIGN AT THE 8 STATIONS

Number of period	Direction of anomaly	Where present
1	Warm	All except Tilsit
2	Warm	All except Glasgow
7	Warm	All 8 stations
9	Cold	All except Kew and Glasgow
15	Cold	All 8 stations
19	Warm	All 8 stations
24	Cold	All except Uppsala and Tilsit
29	Warm	All except Vienna and Glasgow
30	Warm	All 8 stations
31	Warm	All except Uppsala
33	Cold	All except Tilsit
34	Cold	All except Glasgow
35	Cold	All 8 stations
55	Warm	All 8 stations
60	Cold	All except Tilsit and Uppsala
62	Warm	All except Potsdam
64	Cold	All except Uppsala
65	Cold	All 8 stations
73	Warm	All except Glasgow and Uppsala

If the anomalies at the 8 stations were unrelated, then the expectation that all eight for a given period should agree in sign is 1/128, and the expectation that seven of the eight should agree is 1/16. Thus when we examine 73 periods we would expect to find perhaps one period in which all eight agree, and four or five periods in which seven of the eight agree. In fact there are seven periods in which all eight agree, and seven others in which seven of the eight agree, so there is a presumption that the agreements are not wholly due to chance. This presumption is increased when we note from Table 3, that the exceptional stations, when seven of the eight agree, are nearly always on the edge of the area considered, namely Glasgow, Uppsala or Tilsit. This feature is explicable if the weather processes which produce the anomalies are usually on a scale not quite big enough to affect the whole of this area in the same way.

If it is accepted that during the period to which most of these data relate (the 80 years, 1851 to 1930) there are significant departures over this area from the two-term harmonic form, we have yet to consider whether these are likely to be permanent, or to arise from the peculiarities of that particular period. The features likely to be permanent at a place should be identifiable by comparing two records for that place for independent periods of years. Such records are not easy to find, but two have been found, namely, for Prague (Fig. 6) and for Halle (Fig. 7), while there are also mainly non-overlapping records for two pairs of stations which are sufficiently close for us to presume that they will be affected alike in any large-scale process. These pairs are Edinburgh and Glasgow (Fig. 8) and Brussels and Kleve (Fig. 9).

When we compare the anomalies for the periods listed in Table 3 with those in Figs. 6 to 9, we find that in most cases the anomalies are repeated in the records for independent periods of years. In particular it is hardly possible to doubt that for the periods listed in Table 4 the anomalies are systematic and permanent departures from the two term harmonic form.

One other deduction can be made from Figs. 6 to 9. If these anomalies are partly non-random in origin, then there should be a positive correlation between the anomalies found from comparable records for non-overlapping periods of years. These correlations are given in Table 5.

TABLE 4. 5-DAY PERIODS IN WHICH THE DIRECTION OF THE TEMPERATURE ANOMALIES IS CONFIRMED ON LARGELY INDEPENDENT DATA

Number of 5-day period	Dates	Direction of anomaly	Remarks
7	31 Jan.-4 Feb.	Warm	Present in every record, though very small in one or two
15	12-16 Mar.	Cold	Present in all except Edinburgh 1795-1850
19	1-5 Apr.	Warm	Present in all except Prague 1775-1839
29	21-25 May	Warm	Present in all except Vienna and Glasgow, where it is zero
30	26-30 May	Warm	Present at all stations
31	31 May-4 June	Warm	Present at all stations except Prague and Uppsala
35	20-24 June	Cold	Present at all stations except Brussels 1831-1860
55	28 Sept.-2 Oct.	Warm	Present at all stations
64	12-16 Nov.	Cold	Present at all stations except Uppsala
65	17-21 Nov.	Cold	Present at all stations

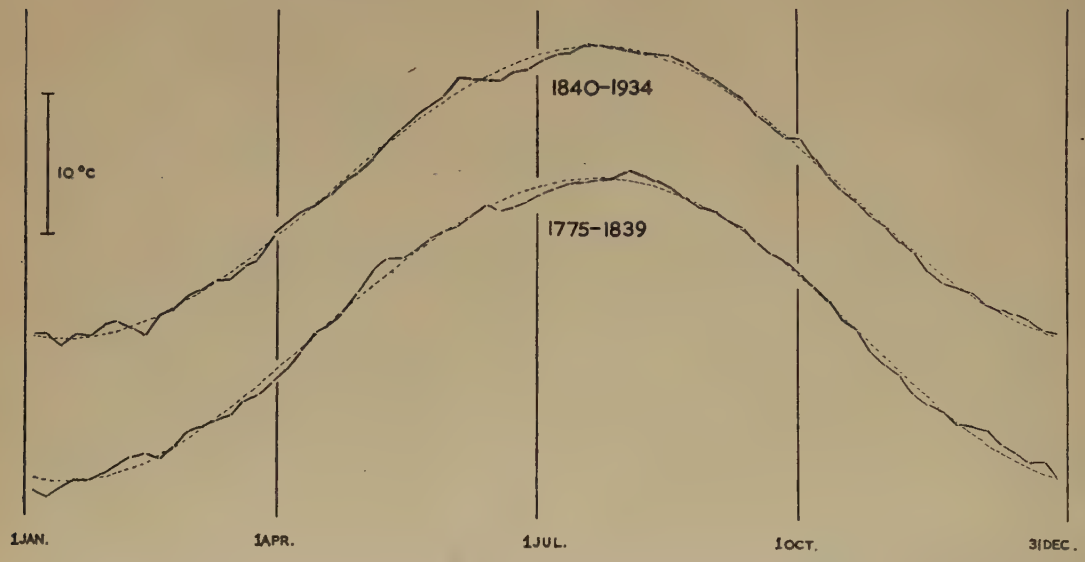

Figure 6. Comparison of temperature anomalies at Prague for the periods 1775-1839 and 1840-1934.

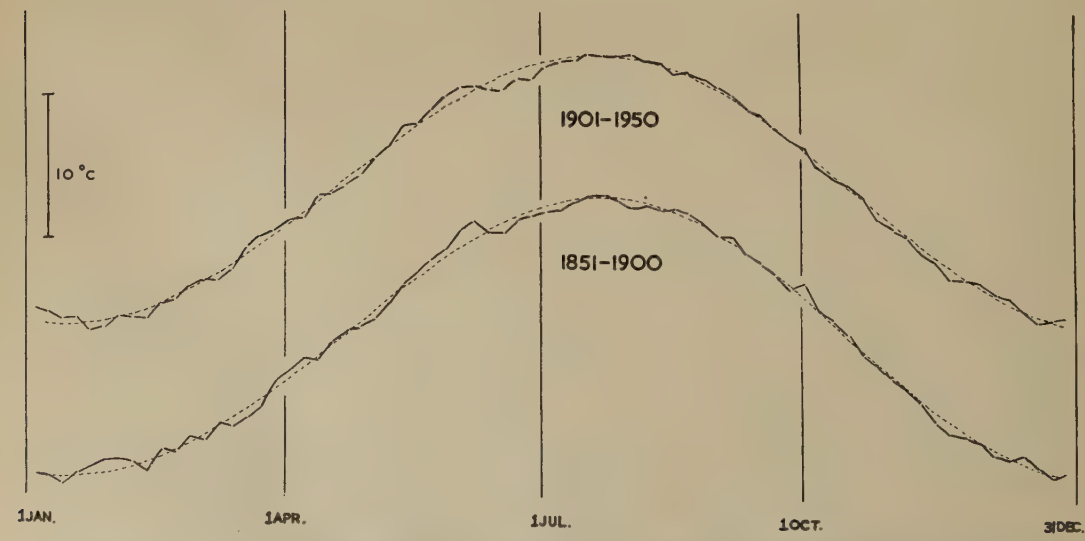

Figure 7. Comparison of temperature anomalies at Halle for the periods 1851-1900 and 1901-1950.

TABLE 5. CORRELATION BETWEEN ANOMALIES FOR WHOLLY OR MAINLY INDEPENDENT PERIODS OF YEARS

First station	Second station	Correlation
Glasgow 1868-1912	Edinburgh 1795-1804 and 1821-1850	0·05
Brussels 1831-1860	Kleve 1851-1930	0·35
Prague 1775-1839	Prague 1840-1934	0·29
Halle 1851-1900	Halle 1901-1950	0·29

The correlation between the anomalies at Edinburgh and Glasgow is evidently not significant; the significance of the other three must be judged in the light of the auto-correlations which exist in the various series of anomalies, the first serial correlations of which are given in Table 6.

Figure 8. Temperature anomalies at Edinburgh 1795-1850 compared with Glasgow 1868-1912.

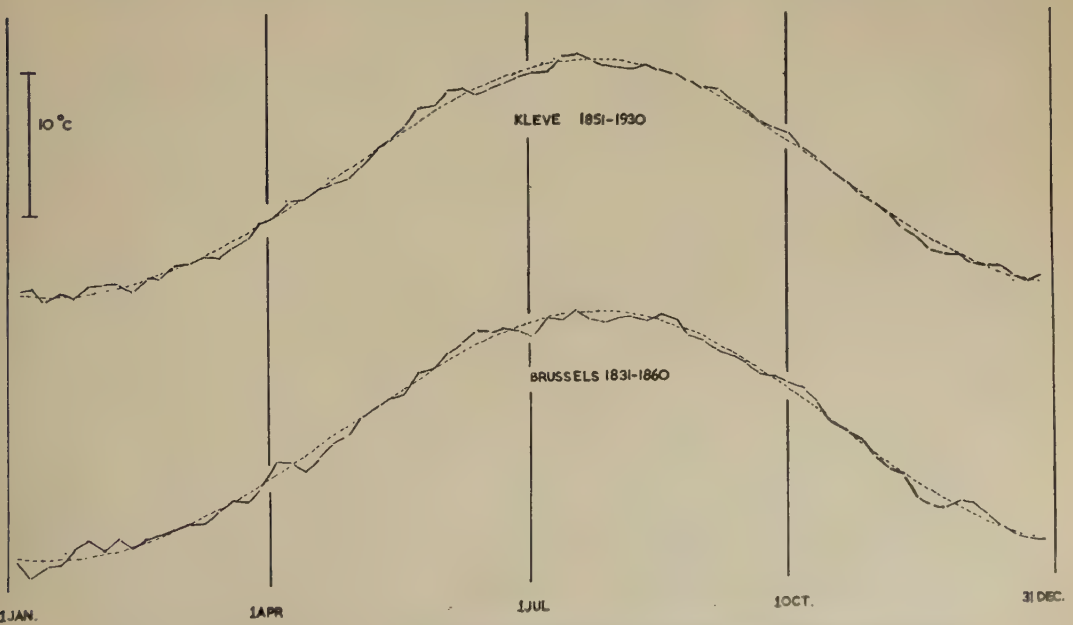

Figure 9. Temperature anomalies at Brussels 1831-1860 compared with Kleve 1851-1930.

TABLE 6. AUTO-CORRELATIONS OF MULTI-YEAR AVERAGES OF 5-DAY MEAN TEMPERATURES

Station	Period	Correlation
Glasgow	1868-1912	0·15
Edinburgh	1795-1804 and 1821-1850	0 46
Kleve	1851-1930	0·56
Brussels	1831-1860	0·49
Prague	1775-1839	0·72
Prague	1840-1934	0·56
Halle	1851-1900	0·45
Halle	1901-1950	0·47

Clearly we cannot treat the correlations given in Table 5 as derived from 73 independent pairs of observations. An estimate following the lines of work by Lewis and McIntosh (1952) gives the 'equivalent number of repetitions' as between 3 and 5, so that the correlation in Table 5 may be judged as based on from 15 to 25 independent pairs of observations. On this basis none of them is significant even at the 10 per cent level. However, the three values for the continental stations are large enough to be suggestive, particularly as the periods of the three comparisons are virtually independent. The absence of association between the anomalies for Edinburgh and Glasgow agrees with the conclusion of Brooks and Mirrlees (1930) who made a similar comparison between temperature anomalies at Kew for the periods 1870-1900 and 1901-1929, so there is a possibility that persistent temperature anomalies may occur over the Continent of Europe without extending to the British Isles.

8. THE SIGNIFICANT TEMPERATURE ANOMALIES

It would be premature to attempt a full-scale discussion of the temperature singularities listed in Table 4. The network of stations was not chosen to show the significance of singularities and there is no guarantee that this list is complete. Nevertheless, it is worth while to consider what has been proved and some possible directions of future investigation. In the first place, the hypothesis that the temperature anomalies are due to local causes independent for each station and period is decisively disproved. On this hypothesis the occurrence of an anomaly of given sign in corresponding periods in each of the 15 records shown in Figs. 5 to 9 should occur only once in 2^{15} investigations, and the number of cases of complete or near complete agreement among the 73 comparisons actually made is far greater than the chance expectation.

Looking at Fig. 5, no one would have been surprised to find the amount of agreement between the anomalies which actually exists had the stations been spaced, say, within a 50-mi radius. What is interesting is that the anomalies should agree to the extent which they do, over an area extending from London to the Baltic. It seems that the processes which produce these anomalies work on a scale rather larger than that of a European country, and approaching the size of the half waves in the general westerly circulation. Further, it looks as though most of the region considered has some kind of climatic homogeneity, since, when the anomaly in a given period has the same sign at all except one station, the exceptional station is usually one on the northern or eastern borders of the area. If we had to outline the area in which this homogeneity might be expected, we should exclude Glasgow, Edinburgh, Uppsala and Tilsit, but include Kew, Potsdam, Vienna, the whole of Germany, and an area to west of the Rhine not covered by present observations.

It is noteworthy that neither the anomalies shown in Figs. 5 to 9 nor those listed as probably significant in Table 4 show any regular arrangement within the year. If we try to represent them by including more harmonics we shall require not one or two extra terms which might have physical significance, but a multitude of high harmonics, each of trifling amplitude. It appears that by taking two harmonics we have extracted all the worth-while information to be obtained by this form of harmonic analysis and that the non-random anomalies are best treated not as the sum of a series of higher harmonics, but as perturbations on the two-term harmonic curve about particular times of year.

9. A particular singularity

The most obvious feature of the annual cycle is the occurrence of positive anomalies in late May and the first few days of June, followed by negative anomalies in late June. The sequence is clearly seen on Fig. 5; for example, in the records for Darmstadt and Potsdam, but it occurs in each of the records within the area described above, and is responsible for three of the widespread anomalies listed in Table 4. The non-casual character of this sequence has been accepted by Bartels (1948) in a paper which otherwise emphasizes the need for caution in accepting singularities as real. The magnitude of the temperature anomaly is largest in the German records, where the positive and succeeding negative anomalies may be nearly 1°C. At Kew the anomalies are similar but a good deal smaller, while they are very small at Edinburgh and Glasgow. At Tilsit and Uppsala the anomalies are smaller and the change from positive to negative anomalies is smoother and less abrupt than at the Central European stations.

This singularity is evidently a phenomenon primarily affecting Central Europe, and becoming less important as we go north and east.

In any part of the world, the climate at a given time of year is the result of several weather types, some warmer, and some colder than the average, and to get an average 5-day mean temperature anomaly of 1°C, which is the figure actually attained in Central Europe, must mean a considerable excess of warm weather types in May followed by an excess of cool types in June. Thus the singularity in central Europe implies a tendency for features of the large-scale circulation to persist in fixed positions during the months of May and June, and if this happens in central Europe, the weather conditions over the British Isles will be materially affected by it. Hence the weather during these months in the British Isles may be influenced by the processes producing a temperature singularity which is most strongly developed in central Europe.

In this case it is easy to suggest a physical cause for the singularity, namely, a monsoonal effect arising from the solar heating of the European land mass, but if we are to forecast the character of the phenomenon from year to year, we should probably look to the area where it is most intense, namely central Europe, or even further east, to Asia where the heating is most intense, rather than to marginal regions such as the British Isles, to find what makes it slight or vigorous, late or early.

10. Discussion

The first part of the investigation shows that over the large area examined the annual temperature variation can be represented for most purposes by two harmonic terms.

The second part of the investigation, which suggests that the annual temperature variation shows small persistent departures from the two-term harmonic form at particular times of the year, is an unexpected by-product of the first. The question, which if any, of these departures will recur in the future cannot be settled beyond doubt from the present

evidence, but there is no doubt that during the period covered by the historical record some of the departures occurred at the same time over a very large area, on the scale of the half-waves of the planetary circulation. We cannot, of course, deduce that disturbances of the planetary circulation are prone to occur in particular areas at particular times of the year, but at least the evidence is consistent with the existence of such disturbances.

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Secular changes of the standing circulation

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Summary

The evidence produced seems to show a secular intensification of the North Indian summer-monsoon low. The effects upon rainfall in North India and the Peninsula are discussed qualitatively. The secular decrease of the mean summer pressures in North India was associated with an increase of winter pressure. Similar pressure changes, with an even larger seasonal amplitude, occurred throughout eastern Europe, Siberia and eastern Asia.

The secular increase in the seasonal pressure oscillation over the interior of Eurasia may have been associated with variations in the wind field which should tend to cause colder winters in the Far East and warmer, shorter winters in Europe. In summer the opposite tendency should prevail. An analysis of records confirms this inference.

An intensification of the standing circulation of the northern hemisphere may have caused an increased transport of warm air into the Arctic and hence contributed to the retreat of the arctic ice. The absence of large-scale standing oscillations in the southern hemisphere, accounts, probably, for the apparent absence of marked climatic change in the Antarctic.

The strength of the standing monsoon-type circulations appears to be in a roughly inverse relation to the strength of the zonal circulation. The resulting interaction could cause a large amplification of small disturbances in the climatic equilibrium.

1. Introduction

In three preceding papers (Kraus 1954, 1955a and 1955b) variations of rainfall regimes have been investigated. It is more difficult to find long series of pressure observations which are internally consistent. Only station-level pressures are suitable. Even then, observation hours and correction procedures are changed from time to time, and this may have a considerable effect on the apparent pressure means.

In cases of doubt the observation series used in the present investigation have been tested wherever possible for internal consistency. This can be done by plotting cumulative residuals, that is cumulative deviations from the average, against the sum of the cumulative residuals for a number of surrounding stations. If the slope of the resulting line changes suddenly, a break in continuity may be expected.

Fluctuations in the mean intensity of continental winter highs and summer lows may be studied conveniently from the changes in the mean January-July pressure difference. The effects of variations in observational procedure, or of small changes in station height upon the annual pressure oscillation, are zero or at least very much smaller than the effects upon absolute values. No trends have been accepted as being physically significant unless confirmed by at least four different series of observations.

Trends are again demonstrated mostly by graphs of cumulative residuals. The merits of this type of representation and the deductions that can be drawn from it have been discussed in an earlier paper (Kraus 1956). The graphs accentuate changes in any time series and they are very useful in demonstrating immediately the existence of trends and the exact time when regimes or trends have changed. The significance of changes can then be tested by other methods.

Deviations have been computed from means of a standard period 1881-1910. A 60-year period would have been preferable, but was not available for a number of stations. The period 1881-1910 was chosen in the first instance because it covers the time of rapid changes in tropical and east coast rainfall regimes in the eighteen-nineties, and because there is a break from 1915 to 1920 in most of the available Russian records.

2. The intensification of the Indian monsoon low

Table 1 shows mean July pressures for four stations in the region of the North Indian monsoon low :

TABLE 1. Mean station-level July pressures in North India (inches)

Station	Height (ft)	Mean (1881-1910) (A)	Mean (1911-1940) (B)	Difference (between means) (B-A)
Lahore (31°34′N, 74°21′E)	702	28·770	28·758	− 0·012
Allahabad (25°28′N, 81°54′E)	309	29·188	29·180	− 0·008
Jaipur (26°55′N, 75°52′E)	1,431	28·105	28·101	− 0·004
Calcutta (22°32′N, 88°24′E)	21	29·509	29·493	− 0·016

It can be seen that the mean July pressure decreased at all four stations. The decrease is not limited to the month of July. For Lahore, mean pressures were also calculated for the combined three summer months : June, July, and August. The linear trend of these summer pressures is $-3{\cdot}5 \times 10^{-4}$ in./year, which Student's t test shows to be statistically significant on the 1 : 1,000 level.

Summer pressure in the North Indian plains appears to be inversely related to rainfall over the same region. To some extent, this negative correlation is used in seasonal forecasting formulae of the Indian Meteorological Department (Savur 1932). However in spite of the secular decrease in the summer pressure, there has been no corresponding increase in the rainfall (Pramanik and Jagannathan 1953). On the other hand, the drop in rainfall volume, which had been found by the author (1955a) to have occurred towards the end of the last century throughout most of the tropics, is also absent from the majority of the North Indian stations, though some which are situated relatively close to the desert – like Jaipur – did in fact experience it. The evidence suggests the factors which influence the summer rainfall regime of North India to be more complex, and perhaps more numerous, than those operating in more equatorial regions. A more detailed analysis, based both on a synoptic approach and on a statistical study of smaller sub-regions with monthly data, might possibly clarify this problem.

3. Summer temperatures of the lower troposphere over North India

Table 2 shows mean July station level pressures for the only two stations in the Himalaya foothills with a sufficiently long series of observations.

TABLE 2. Mean station-level July pressures in the Himalaya foothills (inches)

Station	Height (ft)	Mean (1881-1910) (A)	Mean (1911-1940) (B)	Difference (between means) (B-A)
Leh (34°9′N, 77°34′E)	11,539	19·611	19·622	+ 0·011
Simla (31°6′N, 77°10′E)	7,232	24·933	24·940	+ 0·007

Simla is a relatively short distance from Lahore and Allahabad, but whilst July station level pressures have fallen at the lowland stations, they have risen at Simla. The same holds for Leh, though whether Leh pressure readings can be considered as homogeneous back to 1881 appears to be uncertain. All that can be said is that as far as the data go they suggest an increase of summer temperatures of the lower troposphere over North India. This would be observed as an increased station-level pressure at higher elevations.

4. Changes of surface pressure gradient and rainfall in Southern India

The decrease in July pressures shown in Table 1 seems to indicate an intensification of the monsoon low in the North India plains. Table 3 suggests that a corresponding pressure decrease probably did not occur in the south of India or in Ceylon.

TABLE 3. Mean station-level July pressures in South India and ceylon (inches)

Station	Height (ft)	Mean (1881-1910) (A)	Mean (1911-1940) (B)	Difference (between means) (B-A)
Bangalore (12°58′N, 77°35′E)	3,021	26·785	26·793	+ 0·008
Cochin (9°58′N, 76°14′E)	9	29·824	29·821	− 0·003
Colombo (6°45′N, 79°52′E)	21	29·767	29·779	+ 0·012

Together, the results of Tables 1 and 3 indicate an increase of the meridional pressure gradient in July over the Indian peninsula. This must have caused an increase of the mean westerly wind component, in conformity with the assumed intensification of the North India monsoon low.

Data produced earlier (1954) indicate that rainfall to the lee of mountain ranges is particularly sensitive to changes in airflow and the resulting Föhn effect. The influence of mountains on tropical rainfall can be observed strikingly in East Africa where regions of heavy excess rainfall and extreme aridity may be found within fifty miles' distance in broken country.

It has been shown (1955a) that the May-July rainfall at Madras, on the eastern or lee side of the Indian Plateau, has tended to decrease radically. The argument of the last paragraph suggests that this may have been associated with the increase of westerly wind-strength. As a higher summer rainfall regime had persisted in Madras throughout the 19th century, it would seem that the summer westerlies may have also been weaker throughout this period.

There was no significant summer rainfall increase on the Indian West Coast corresponding to the decrease in the East. In the earlier paper (1955) it had been argued that the very widespread decrease of tropical rainfall at the end of the last century must have been associated with a decrease in the intensity and moisture content of the trade winds. The absence of significant trends in the rainfall regime of the Indian West Coast suggests that the effect of decreased moisture content in the air mass which originated in the trades of the Southern Indian Ocean, and the effect of the increased on-shore wind component, have perhaps tended to balance each other to some extent.

The opposite holds for the autumn and winter rains of Madras. These were shown in the earlier paper to have increased considerably. The explanation may be an increase in the strength of the on-shore NE monsoon which in this case may not have been balanced

by a lower moisture content. An increase of the NE monsoon should be expected from the observed intensification of the Asiatic winter anticyclone which will be discussed below.

The meridional pressure gradient over India may be related not only to the distribution of monsoon rainfall, but also to the date of its onset along the West Coast. Ramdas, Jagannathan and Gopal Rao (1954) have found a negative correlation between this date and the Cochin-Jaipur pressure difference in April. From the data presented by these authors it can also be deduced that the monsoon established itself at the Travancore-Cochin coast on the average 4·5 days earlier during the period 1891-1910 than during the period 1911-1930. This would suggest an increasing meridional pressure gradient. Although in itself hardly significant, it lends some further support to the present argument.

5. Changes of the Indian monsoon oscillation

Whilst the summer low over North India has tended to deepen, the winter pressures have tended to increase. In fact Table 4 shows the rise of January pressures to be considerably more pronounced than the fall of July pressures.*

TABLE 4. Mean station-level January pressures over North India (inches)

Station	Mean (1881-1910) (A)	Mean (1911-1940) (B)	Difference (between means) (B-A)
Lahore	29·312	29·327	+ 0·015
Allahabad	29·723	29·735	+ 0·012
Jaipur	28·570	28·586	+ 0·016
Calcutta	30·008	30·017	+ 0·009

Tables 1 and 4 together indicate that the Indian monsoon oscillation, that is the January-July pressure difference, has increased in amplitude between periods (A) and (B). In other words, there must have been an increasing net horizontal convergence of the air flow in autumn and a correspondingly greater divergence in spring over North India.

It is of considerable interest to study the time characteristics of this change. This is done most economically by a plot of cumulative residuals of the January-July pressure difference as shown in Fig. 1. The (1881-1910) mean of the pressure difference $\overline{\Delta p}$ is indicated for each station separately in the figure. It may be recalled from the earlier papers that if Y_m and Y_{m+n} are the values of the cumulative residuals for the years m and $m + n$, the mean of the intervening years differs from $\overline{\Delta p}$ by $(Y_{m+n} - Y_m)/n$, i.e. by the slope of the graph.

The most striking feature of the figure is perhaps the non-linearity of the change. It illustrates the danger of deducing long-term climatic fluctuations from comparison of arbitrary 30-year means. For example, the mean January-July pressure amplitude at Calcutta increased from 0·499 in. for the years 1881-1910 to 0·524 in. for 1911-1940. Fig. 1 shows immediately that no comparable increase occurred between the 30-year periods 1871-1900 and 1901-1930. A comparison of the periods 1861-1890 and 1891-1920 would actually suggest a decrease of the pressure amplitude. In fact, it is sometimes possible to deduce almost any sort of change from a suitable choice of 30-year means.

* Since this paper was prepared a new article by Pramanik and Jagannathan (1955) has come to hand. From the data given by these authors it would appear that the rising trend of January pressures over the period (1901-1950) was limited to the Indus valley plains, as shown by the records of Jacobabad, Multan and Lahore. It did not occur further west or south. The 1901-1950 July pressure trends agree with the deductions presented above.

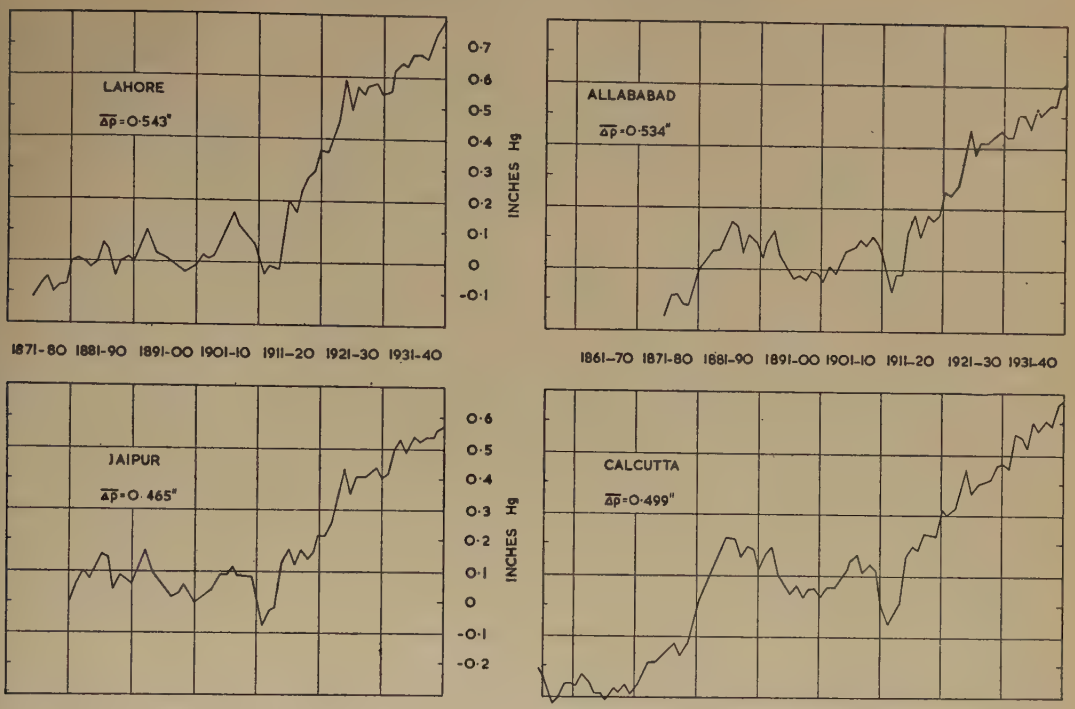

Figure 1. Cumulative deviations of the January-July pressure amplitude (Δp) from its 1881-1910 average ($\overline{\Delta p}$).

What Fig. 1 does prove, with little apparent doubt, is a sudden increase of the mean January-July pressure amplitude or monsoon oscillations after 1912. The years 1885-1911 were mostly characterized by relatively low values of the January-July pressure amplitude. Before 1885, the graphs for Allahabad and Calcutta suggest the possibility of another period of comparatively vigorous monsoon oscillation. This may have begun in the eighteen-fifties – if analogy with the temperature data, discussed below, provides any guide.

6. Changes of the main Asiatic pressure oscillation

The secular changes of the Indian monsoon oscillation appear to have been paralleled on a much larger scale throughout eastern Europe, Siberia and eastern Asia. This is demonstrated in Fig. 2. The graphs of this figure were calculated from basic data which give the pressure in millimetres. An inch scale is shown however on the outer side of the figure. The fluctuation which occurred north of the great mountain ranges was very much larger than that deduced for northern India, and the vertical scale of Fig. 2 is reduced to one-half that of Fig. 1. There is a considerable similarity between the curves. In particular, the graphs for the northern Asiatic stations all show the same steep rise after 1912, as do, to a lesser extent, the graphs of the North Indian stations. It is concluded that the increase in the amplitude of the pressure oscillation was a widespread phenomenon of considerable magnitude.

The increase of the seasonal pressure amplitude appears to have been mainly due to an intensification, or a more frequent occurrence, of the anticyclones which are a regular feature of the winter weather over most of Eurasia. The decrease in summer pressure is less pronounced and can be observed only in the interior of the continent. Tables 5 and 6 show this for the four stations represented in Fig. 2. A marked increase of January pressure can be shown to have occurred not only on these four stations, but also on all other Russian stations listed in *World Weather Records*.

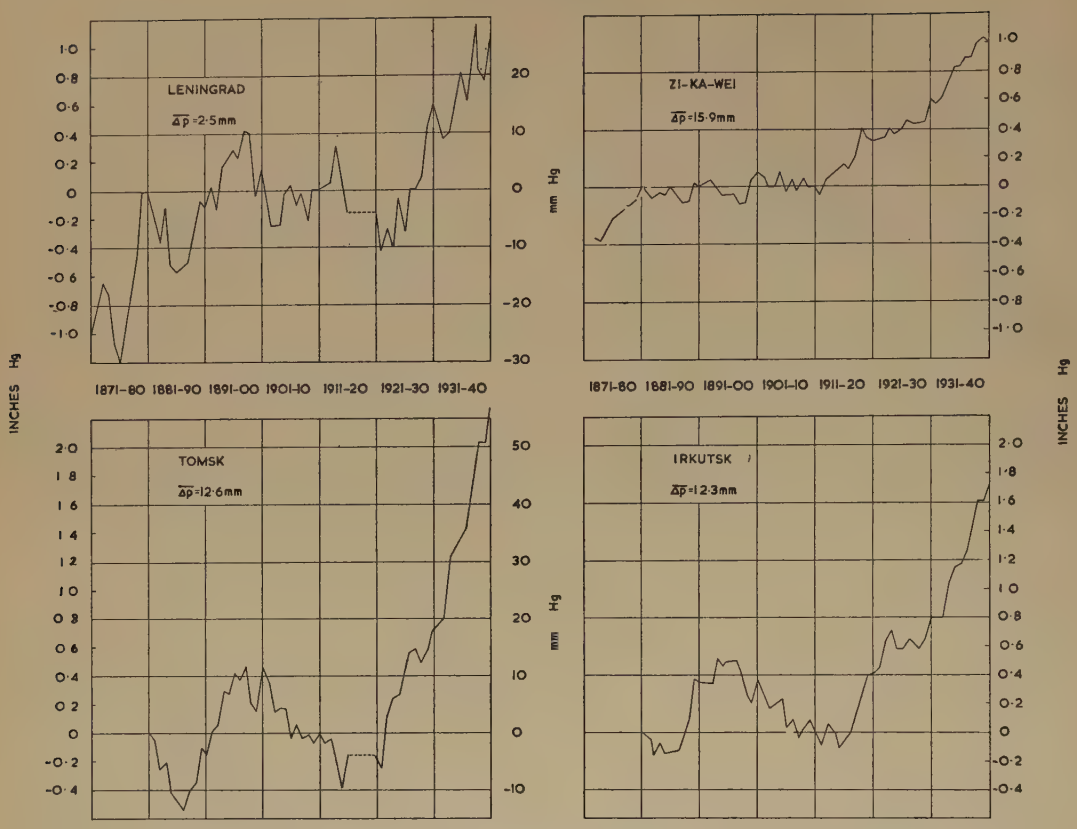

Figure 2. Cumulative deviations of the January-July pressure amplitude (Δp) from its 1881-1910 average ($\overline{\Delta p}$).

TABLE 5. MEAN STATION-LEVEL PRESSURES FOR JANUARY (INCHES)

Station	Height (ft)	Mean (1881-1910) (A)	Mean (1911-1940) (B)	Difference (between means) (B-A)
Leningrad (59°58′N, 30°17′E)	26	29·917	29·972	+ 0·055
Tomsk (56°26′N, 84°58′E)	404	29·815	29·882	+ 0·067
Irkutsk (51°16′N, 104°19′E)	1,532	28·630	28·681	+ 0·051
Zi-ka-wei (31°11′N, 121°26′E)	23	30·272	30·331	+ 0·059

TABLE 6. MEAN STATION-LEVEL PRESSURES FOR JULY (INCHES)

Station	Height (ft)	Mean (1881-1910) (A)	Mean (1911-1940) (B)	Difference (between means) (B-A)
Leningrad (59°58′N, 30°17′E)	26	29·819	29·823	+ 0·012
Tomsk (56°26′N, 84°58′E)	404	29·319	29·307	− 0·012
Irkutsk (51°16′N, 104°19′E)	1,532	28·146	28·130	− 0·016
Zi-ka-wei (31°11′N, 121°26′E)	23	29·646	29·673	+ 0·027

Comparison of Tables 4 and 5 shows that the January pressure increase in Siberia was about four or five times that of North India.

It might be noted that the series of pressure records from Zi-ka-wei is almost certainly not consistent internally. A change in observational or recording procedure appears to have occurred about 1930. All pressure readings thereafter were higher than before. There is however no reason to doubt the veracity of the January-July pressure difference, which is the basis of Fig. 2.

7. The effect of circulation changes upon temperature in East Asia

An increase of the mean winter pressure over the continent is likely to have been associated with more frequent or more vigorous north winds over the China coast and Japan. It might be expected that this would lead to colder winters. This assumption appears to be confirmed by the data of Table 7, which show that January temperatures in East Asia did in fact decrease, or at least they did not increase as happened in Europe, the Arctic and elsewhere.

In summer the opposite should hold. The slight mean pressure decrease inland should lead to somewhat stronger or more frequent south winds and hence to a corresponding increase in temperature. Table 8 indicates that this, too, seems to have been realized.

TABLE 7. Mean January temperatures (Centigrade) in East Asia

Station	Mean (1881-1910) (A)	Mean (1911-1940) (B)	Difference (between means) (B-A)
Hongkong (22°18′N, 114°10′E)	16·0*	15·2	− 0·8
Zi-ka-wei (31°11′N, 121°25′E)	3·6	3·2	− 0·4
Kioto (35°1′N, 135°44′E)	2·6	2·4†	− 0·2
Tokyo (35°41′N, 139°46′E)	3·0	3·0†	0·0

TABLE 8. Mean July temperatures (Centigrade) in East Asia

Station	Mean (1881-1910) (A)	Mean (1911-1940) (B)	Difference (between means) (B-A)
Hongkong (22°18′N. 114°10′E)	27·7*	27·9	+ 0·2
Zi-ka-wei (31°11′N, 121°25′E)	26·8	27·3	+ 0·5
Kioto (35°1′N, 135°44′E)	24·9	25·8†	+ 0·9
Tokyo (35°41′N, 139°46′E)	23·8	24·7†	+ 0·9

* Mean for years 1884-1910. † Mean for years 1911-1936.

Fig. 3 illustrates the change of the July-January temperature amplitude $\overline{\Delta T}$ in time. The similarity of the temperature graphs with the graphs of the pressure amplitude is immediately evident. In particular, all the diagrams show a marked steepening of the curve – that is an increase of the amplitude – after about 1911.

8. A comparison of secular temperature changes in Europe and East Asia

The temperature change in Europe may be expected to be opposite to that in East Asia. High winter pressures over Russia and Siberia should be associated with south winds or east winds over parts of Europe. In January these winds can bring very cold weather

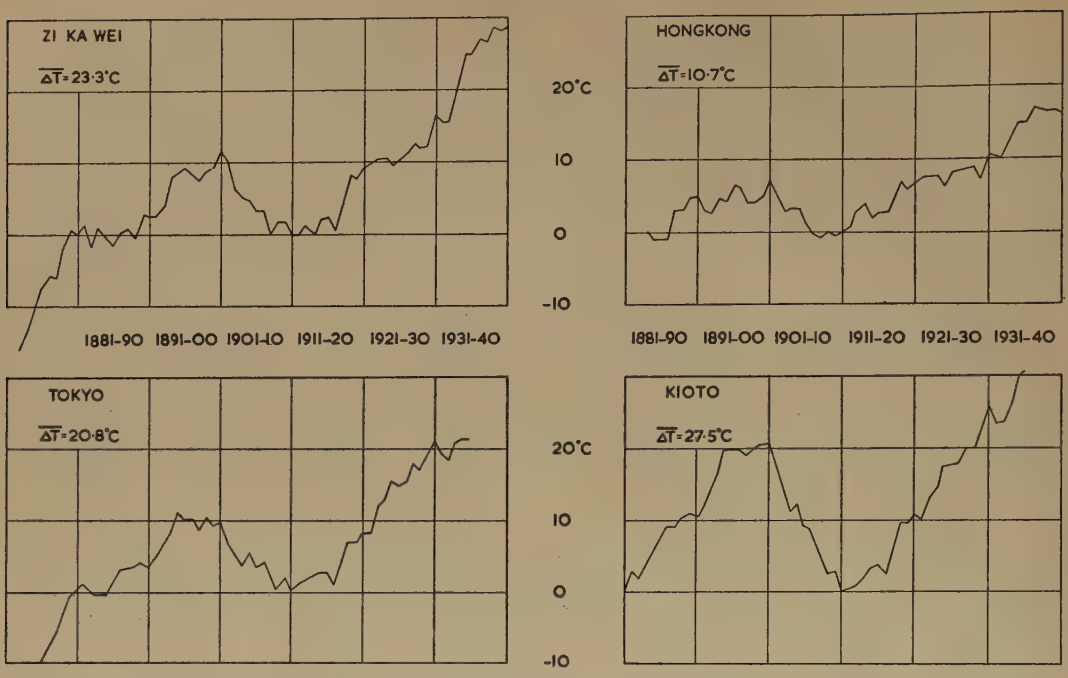

Figure 3. Cumulative deviations of the July-January temperature amplitude (ΔT) from its 1881-1910 average ($\overline{\Delta T}$).

to the Balkans, the Danube basin and the East European plains. However, later towards spring, they may cause a tendency towards higher temperatures, and the same will apply at all times to the lee of the Alps. On the other hand, low summer pressures over the interior of the Eurasian continent will tend to bring cold air from the north into Europe.

The warming up of the European winter has been studied by many authors. A list of references is given by Ahlmann (1953). In order to provide a comparison with conditions in East Asia, data for four stations are given in Tables 9 and 10.

TABLE 9. Mean January temperatures (Centigrade) in Europe

Station	1851-1880 (A)	1881-1910 (B)	1911-1940 (C)	Difference (B-A)	Difference (C-B)
Vienna	− 1·38	− 1·82	− 0·08	− 0·44	+ 1·74
Zurich	− 1·07*	− 1·53	0·19	− 0·46	+ 1·72
Utrecht	2·32	1·71	2·72	− 0·61	+ 1·01
Uppsala	− 3·77†	− 3·95	− 3·65	− 0·18	+ 1·30

TABLE 10. Mean July temperatures (Centigrade) in Europe

Station	1851-1880 (A)	1881-1910 (B)	1911-1940 (C)	Difference (B-A)	Difference (C-B)
Vienna	19·59	19·27	18·77	− 0·32	− 0·50
Zurich	18·86*	17·93	17·59	− 0·93	− 0·34
Utrecht	18·95	18·41	18·51	− 0·54	+ 0·10
Uppsala	16·53†	16·14	17·36	− 0·39	+ 1·22

* Mean for years 1864-1880. † Mean for years 1855-1880.

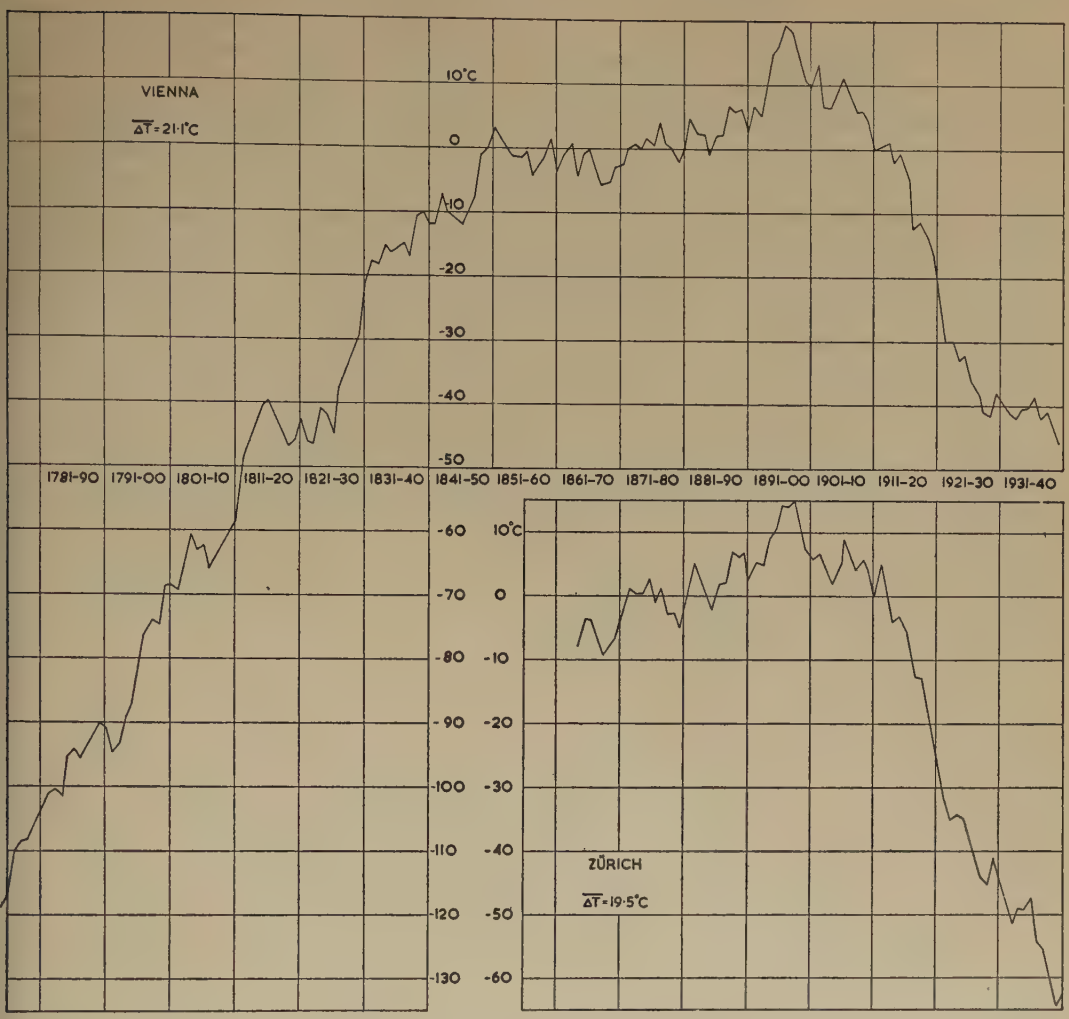

Figure 4. Cumulative deviations of the July-January temperature amplitude (ΔT) from its 1881-1910 average ($\overline{\Delta T}$).

Comparison of Tables 9 and 10 with Tables 7 and 8 suggests in fact the expected reversal of the temperature change on the eastern and western side of the Eurasian land mass.

It is of some interest that the European winters of the period 1881-1910 appear to have been cooler on the average than those of the preceding 30 years. This may possibly be related to the time characteristics of the pressure change. Figs. 1 and 2 show a period of comparatively very weak seasonal pressure oscillations between 1898 and 1911. The associated low values of the temperature amplitude in East Asia are indicated by Fig. 3.

The time change of the temperature amplitude in the west is perhaps best discussed by reference to Fig. 4. The graphs for both Zurich and Vienna show a steeply descending slope after 1911 which signifies a period of warm winters and comparatively cool summers. The same period was characterized by comparatively cold winters and warm summers in East Asia and by a relatively large January-July pressure amplitude in North India, Siberia and Russia. This would seem to confirm the present argument that secular temperature variations in Europe and Asia are closely associated with the position and intensity of the

quasi-stationary winter highs and summer lows of the Eurasian continent. In particular, periods of relatively large seasonal temperature amplitudes in the east should be associated with periods of relatively small amplitude in the west and vice versa.

Willett (1950) has published charts of 20-year mean changes of temperature centred on 1910 and 1930. These seem to agree reasonably well with the scheme presented here, considering that the most rapid change occurred about 1911, that is in the middle of Willett's first period. Willett suggested also that the main cause for temperature change was intensification of the continental winter anticyclones, both over Eurasia and North America. In the present study, the North American data were not used, because they appear to be influenced by a change in the intensity or wavelength of lee waves to the East of the Rocky Mountains. To separate these from the effect of changes in the North American ' monsoon ' circulation would require a more complex and refined analysis.

The graph for Vienna shows a relatively large seasonal temperature amplitude before 1850 all through the first half of the nineteenth and the latter part of the eighteenth century. During this period winters in Vienna tended to be cold and summers were comparatively warm. Manley's (1946) and Wallén's (1953) investigations show that the same happened in other parts of Europe. This would indicate a more ' continental ' character of the European weather. If the argument at the beginning of this section is correct, this should have been associated with relatively warmer winters and colder summers during the same period over the greater part of eastern Asia.

9. The effect of arctic conditions on the temperature amplitude

Comparison of Table 7 with Table 9 shows the winter temperature decrease in East Asia to have been much smaller than the corresponding temperature increase in Europe. This is probably due to the warming of the Arctic. Although winter north winds may have been stronger or more frequent over East Asia during the second period, they came from a warmer Arctic. The two effects should have tended to balance each other to some extent.

The opposite holds for July. The temperature increase in East Asia is comparatively large. The corresponding decrease in Europe is only in evidence at the southern stations. July temperatures actually increased in Utrecht and even more in Uppsala. This again is probably related to the warming up of the Arctic, which might cause North winds over Northern Europe to be relatively warmer.

10. Glaciation changes in the Arctic and the Antarctic

There is a considerable discrepancy between the effects of the recent climatic fluctuation in the Arctic and the Antarctic. The large retreat of glaciers and sea ice in the north apparently did not occur in the south; at least it must have been much less pronounced.

The present investigations suggest an association between the warming of the Arctic and the increase of the standing Asiatic circulation. Relatively high winter pressures over Asia would tend to produce an increased transport of warm air into the Arctic on their western side; lower summer pressure should have the same effect in the east. It has been shown above that the amplitude of the monsoon oscillation was comparatively large between 1911 and 1940. Another period of comparatively high seasonal pressure oscillations appears to have occurred before 1880 as evidenced by the pressure curves for Allahabad, Calcutta and Irkutsk. The temperature curve for Vienna suggests that this first period may have begun about 1850. The most active retreat of many glaciers occurred roughly

during the same two periods (Ahlmann 1953) and may have been causally connected with the circulation change. If this inference is correct we might expect glaciers in Kamchatka for example to have reacted differently from glaciers in the Alps or in Scandinavia. The effects of circulation changes on glaciation are however necessarily complex. The operative factor is not only temperature advection, but also differences in the precipitation pattern, in drift-ice limits, in cloudiness, etc.

Willett has already drawn attention to the fact that there are no large standing circulations in the southern hemisphere, and that this seems to be a possible explanation for the non-existing or less pronounced change of ice conditions in the Antarctic.

11. The inter-relation between zonal winds and the standing circulation

It has been found (Kraus 1954) that strong developments of a westerly sub-tropical jet stream over Australia between 1949 and 1954 were associated with low monsoon rainfall in Queensland and vice versa. This appears to be a fairly general phenomenon. The atmosphere can be considered as a circular vortex about the polar axis with superimposed disturbances. The dynamic stability of this vortex will be increased by strong upper westerlies close to the equator. This should tend to hamper the development of monsoon-type disturbances.

If this can be confirmed, it would allow us to stipulate a mechanism of climatic change. During the cooler period prior to 1850 the atmosphere of the northern hemisphere was probably characterized by less active monsoon circulations and by a more zonal type of weather. That means that conditions resembled more closely those of the southern hemisphere than they do today. The same, in an even more pronounced form, may have occurred during the cold, wet, ' sub-Atlantic ' period which lasted from about 1,000 B.C. to 300 A.D. (Brooks 1951) and perhaps also during the ice age. A slight decrease of the zonal circulation would lead to a stronger development of the continental winter highs and summer lows, which in turn would cause a greater northwards transport of warm air. The process is cumulative as the warming of the Arctic and the resulting retreat of the ice would probably tend to reduce the zonal circulation still further, and so on.

Climatic fluctuations in the tropics are probably affected by a different type of instability, as described earlier. A weakening of the sub-tropical easterlies will reduce the amount of evaporation from the tropical oceans. This will cause a reduction of rainfall in the inter-tropical convergence zone. The associated decrease of the heat of condensation will tend to decrease still further the amount of energy which is made available to drive the trades, and so on.

There are hence possibly two distinct processes in the tropics and in higher latitudes which should both tend to amplify any small disturbance of the climatic equilibrium. Over the past hundred years or so, both processes appear to have been associated with a reduction of the zonal circulation, that is with a decrease of both the low-level easterlies in the tropics, and the westerlies at upper levels and at higher latitudes. This should be related to the reported decrease of tropical rainfall and the warming of the Arctic. On the other hand, an increase of the zonal wind system should produce a tendency towards glaciation in high latitudes and a pluvial age in the tropics.

Both the processes discussed above would be affected primarily by small meridional changes of the radiation balance. For this reason and from consideration of the requirement of an angular momentum balance along the surface, we should expect both processes to have influenced the climate more or less concurrently. However they cannot be expected to have acted at the same rate and this may account partly for the non-uniform, pulsating character of the observed climatic change.

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The decay of the ice-nucleating properties of silver iodide released from a mountain top

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Summary

Silver-iodide freezing nuclei, together with a tracer of zinc-sulphide particles, were released from a mountain top and detected in an aircraft. The ice-nucleating properties of the silver iodide decayed in the daytime but not at night. The ambient temperature did not appear to affect the daytime rate of decay. The top of the smoke plume rose no higher than in previous experiments when it was released from a flat plain.

1. Introduction

Silver-iodide freezing nuclei have been widely used in attempts to stimulate the formation of rain. They have usually been released from generators on the ground, in the hope that they will rise into suitable supercooled clouds without losing their properties.

The durability of the nuclei has been the subject of many investigations in the laboratory, the results of which do not agree well. For example Reynolds, Hume, Vonnegut and Schaeffer (1951) reported that the nuclei rapidly lost their effectiveness when exposed to ultra-violet radiation. This was denied by Vonnegut and Neubauer (1951). The rate of decay of ice-nucleating properties of silver iodide has been reported by Bolton and Qureshi (1954) to be a function of the ambient temperature. This has been denied by Warner and Bigg (1956).

Experiments by Smith and Heffernan (1954) and by Smith, Heffernan and Seely (1955) showed that in the free atmosphere the ice-nucleating properties of silver iodide decayed at a rate which depended on the method of generation of the nuclei. The cause of the decay was not investigated. In these experiments the nuclei were tracked for up to two hours after release; during this time they did not rise to altitudes where the temperature was low enough for them to initiate the formation of ice crystals in suitable clouds.

The present paper describes a further series of measurements in which silver-iodide freezing nuclei, together with a tracer aerosol of zinc sulphide, were released from a mountain top and detected in an aircraft. The ambient temperatures were lower than those prevailing in the previous experiments and measurements were made both by day and by night. The intention was to investigate the effect of daylight and temperature on the rate of decay of the ice-nucleating properties of the silver iodide, and to determine whether the higher point of release caused the nuclei to rise to higher altitudes.

2. Location of trials

In the series of trials described in the present paper the smoke generators were mounted on the top of Mt. Wellington, near Hobart, Tasmania, Australia. The altitude is 1300 m above sea level. The trials took place between 5 July and 5 August 1955. This mountain top was selected because, in winter, it best fulfills the requirements of height, low temperature, easy access and reasonable proximity to Sydney. The terrain to the west of Mt. Wellington is mountainous, with peaks of similar height. To the north and south the hills are lower, to the east and north-east the ground is comparatively low and flat, while to the south-east there is water. Since the top of the smoke plume was usually no higher

than the mountain top (see Section 5) daytime flights through the smoke could only take place with winds from south through west to north. At night, owing to the necessity for increased terrain clearance for reasons of aircraft safety, flights could only take place with westerly or north-westerly winds. The generators were mounted at the highest point of the mountain, close to the edge of a high cliff, where the ground falls to the east. With the prevailing westerly winds it might have been better to have placed the generators further to the westward in order to avoid the downdraught over the cliff. However, this was the only accessible position.

3. Apparatus

An acetone solution of silver iodide was burnt in either a hydrogen or a kerosene flame. The hydrogen burner was of the type described by Vonnegut (1949) and was identical with that used in the previous trials (Smith *et al.* 1955). It consumed per hour 200 g of silver iodide. In the kerosene burner previously used (Smith *et al.* 1955) an acetone solution of silver iodide was atomized into the flame of a large kerosene blowlamp. This type of burner could not be used on the top of Mt. Wellington because it would not burn reliably in the strong winds sometimes experienced there. Accordingly a burner was constructed in which the kerosene flame and the spray of silver-iodide solution were enclosed in a flame tube. The consumption of silver iodide was 640 g/hr. The temperature recorded by a thermojunction of 0·1 mm platinum-platinum/rhodium wire in the outlet of the flame tube was 1,400°C compared with 1,440°C in the flame of the kerosene blowlamp previously used.

The zinc-sulphide aerosol was dispensed by a dry duster as described by Braham, Seely and Crozier (1952). Its output was $3{\cdot}4 \times 10^{13}$ particles/hr. The powder used was Fluorescent No. 2210 of the New Jersey Zinc Corporation. Most of the particles were between one and two microns, with a terminal velocity of about 1 m/hr. Thus their trajectory should have been, for all practical purposes, identical with that of the silver-iodide particles whose terminal velocity is negligible.

The silver-iodide freezing nuclei were detected by a cold chamber of 100 l capacity (as described by Smith and Heffernan 1954), installed in a Dakota aircraft. This was filled with air from outside the aircraft, and cooled at 4°C/min, water vapour being added to form a fog. The concentration of freezing nuclei was derived from the observed concentration of ice crystals at − 17°C (Smith and Heffernan 1954).

The zinc sulphide was detected by an impactor, of the improved type described by Crozier and Seely (1955). The particles impinged on and adhered to an adhesive strip, which moved at either 0·046 or 0·139 mm/sec past the entry orifice, of 6·3 mm by 1·27 mm. The particles which were illuminated by ultra-violet light could be observed at the point and time of impact by their fluorescence. Quantitative determination of the distribution of particles on the adhesive strip took place subsequently on the ground. Timing marks identified the position of particles on the recording strip with the position of the aircraft when they were collected. The air flow through the impactor was 75 l/min.

Thus a continuous record was provided of the concentration of zinc sulphide in the air through which the aircraft flew, while the freezing-nucleus concentration was measured at selected points.

4. Description of trials

A silver-iodide smoke-generator and the zinc-sulphide dispenser were set in operation simultaneously. After a suitable interval the aircraft started flying through the smoke at right angles to the wind. On some flights the altitude was varied from above the top of the smoke plume to the bottom (or as near to the bottom as was practicable) but on others it was maintained constant.

On each pass through the smoke, the operator of the zinc-sulphide impactor reported when the aircraft was in the smoke plume and, when its intensity appeared to be at a maximum, an air sample was admitted into the cold chamber. The aircraft's altitude was then maintained constant while the sample was analysed for silver-iodide freezing-nucleus content. Several measurements were made at intervals of about 12 min at the same distance from the smoke generator.

After the flight, the zinc-sulphide aerosol-concentration, at the times and places where the silver-iodide samples were collected, was computed from the particle density on the impactor record strip. Thus in each series of measurements, at the same distance from the smoke generators, the ratio of the concentrations of silver-iodide freezing nuclei to zinc-sulphide particles was determined several times. Inspection of these measurements suggested that they were distributed exponentially rather than linearly. Therefore the geometric mean of the series of measurements of this ratio was multiplied by the flux of zinc-sulphide particles to give an estimate of the flux of effective freezing nuclei.

On all flights it was established, before and after the passes through the smoke, that no ice crystals formed at temperatures above — 20°C in air samples taken outside but level with the smoke plume.

5. Results of trials

Ten flights were made, in which 11 measurements of freezing-nucleus flux were made with the kerosene silver-iodide burner by day, 8 measurements using the same burner by night, and 4 measurements using the hydrogen silver-iodide burner by night. The results, and the conditions in which they were obtained, are summarized in Table 1, in which each line refers to a series of several measurements. The distance given in

TABLE 1. Summary of results

Date (1955)	Silver-iodide burner (type)	Time	Distance (km)	Wind speed (m/s)	Exposure time (min)	Cloud		Flux of freezing nuclei (per gram AgI)	Plume width (km)	Top of plume (m)
5 July	Kerosene	Day	10	8	21	3/8Ac	1/8Sc	7×10^{13}	3	1300-1400
			34	8	71	3/8Ac	1/8Sc	3×10^{13}	7	> 1800
12 July			10	13	13	3/8Ac	3/8Sc	3×10^{13}	5	
			34	13	44	3/8Ac	3/8Sc	$1{\cdot}3 \times 10^{12}$	8	1200-1500
17 July			10	10	17	1/8Sc		$1{\cdot}8 \times 10^{14}$	3	
			31	10	52	1/8Sc		5×10^{13}	7	1500-1700
18 July			31	15	34	1/8Ci	5/8CuSc	$1{\cdot}8 \times 10^{13}$	9	
20 July			8	5	27	5/8Sc		7×10^{13}	3	1200-1400
			23	5	77	5/8Sc		$1{\cdot}1 \times 10^{12}$	6	
24 July			4	6	11	7/8CiSt	1/8Sc	5×10^{13}	4	1200-1400
			29	6	80	7/8CiSt	1/8Sc	$1{\cdot}8 \times 10^{13}$	9	
25 July	Kerosene	Night	10	4	42	Nil		5×10^{14}	5	1400-1500
			19	4	79	Nil		9×10^{14}	7	
			35	4	146	Nil		7×10^{14}	11	
26 July			10	7	24	1/8CiSt		4×10^{14}	5	
			26	7	62	1/8CiSt		4×10^{14}	8	
			37	7	88	1/8CiSt		6×10^{14}	8	
5 Aug.			10	5	33	Nil		4×10^{14}	5	
			18	5	60	Nil		5×10^{14}	6	
3 Aug.	Hydrogen	Night	10	10	17	2/8Sc		3×10^{14}	3	
			35	10	58	2/8Sc		$1{\cdot}7 \times 10^{14}$	11	
			47	10	78	2/8Sc		3×10^{14}	13	
5 Aug.			10	4	42	Nil		2×10^{15}	9	

column 4 is that between the smoke generators and the point where the silver-iodide sample was taken; it did not vary significantly between measurements in one series. The wind speed was measured by pilot balloons released at half-hourly intervals at Cambridge airport, 20 km east of the smoke generators, and confirmed by drift measurements in the aircraft. The ' wind speed ' given is the mean of the measurements between the time the smoke generators were set in operation and the time of the last measurement, and from the top to the bottom of the smoke plume. These limits of altitude were sometimes determined experimentally but on other occasions were arbitrarily assumed on the basis of the sounding and previous experience. Errors thus introduced are not thought to be significant as there was little wind shear. There was also little variation of wind with time during any one experiment. The plume widths given in column 9 are deduced from the zinc-sulphide record, and are the mean of the values obtained on the several passes of the series. They are arbitrarily taken to include all but a few of the zinc-sulphide particles. On six occasions two figures are given in column 10, under the heading ' Top of plume.' Zinz sulphide was detected in a pass or passes at the lower altitude given, but none was detected in a long pass or passes at the higher altitude.

The individual measurements of silver-iodide freezing nuclei and zinc-sulphide concentration are given in Tables 2 to 4.

Temperature soundings on the days of the flights are given in Table 5, the lapse rate being uniform between the altitudes given. Altitudes are above sea level. The temperature measurements were made in the aircraft at the beginning of each flight, except on 20 July, when the aircraft thermometer was unserviceable and the figures are quoted from the radiosonde sounding at Cambridge airport at 1300 hr LT, the flight starting at 1345 hr. Temperature inversions were usually present at or just above the altitude of the mountain top. On 3 August the sounding extended to 1200 m, which was the height of the base of a layer of stratocumulus. There was probably an inversion at the top of the layer, the altitude of which was estimated as 1,500 m.

TABLE 2. INDIVIDUAL MEASUREMENTS : kerosene burner (day)

Date (1955)	Distance (km)	A = AgI nuclei/l Z = ZnS particles/l								
5 July	10	A	3,000	10,000	3,000	10,000	50	10		
		Z	4	3	3	0·2	0·08	0·2		
	34	A	20	3,000	500					
		Z	0·2	1·0	0·9					
12 July	10	A	4,000	300	300	30	15			
		Z	0·7	0·08	0·8	0·4	0·15			
	34	A	10	5	4	1				
		Z	0·05	0·2	0·5	0·17				
17 July	10	A	300	2,000	500					
		Z	0·2	0·2	0·15					
	31	A	20	30	30	30	15	30	200	300
		Z	0·05	0·07	0·07	0·05	0·05	0·02	0·07	0·05
18 July	31	A	100	50	30	2				
		Z	0·15	0·07	0·12	0·02				
20 July	8	A	1,000	50	200	4,000				
		Z	0·5	0·5	0·15	0·5				
	23	A	2	6	5	1				
		Z	0·2	0·3	0·10	0·05				
24 July	4	A	3,000	10,000	30,000	3,000	50			
		Z	2	5	15	3	0·2			
	29	A	10	300	300	300	500			
		Z	0·07	0·3	1·4	1·5	0·7			

TABLE 3. INDIVIDUAL RESULTS : kerosene burner (night)

Date (1955)	Distance (km)	A = AgI nuclei/l Z = ZnS particles/l					
25 July	10	A	3,000	300	5,000	4,000	2,000
		Z	0·15	0·15	0·4	0·2	0·4
	19	A	3,000	5,000	7,000	700	5,000
		Z	0·15	0·4	0·3	0·07	0·19
	35	A	2,000	3,000	500	700	
		Z	0·07	0·19	0·12	0·07	
26 July	10	A	2,000	10,000	10,000	10,000	
		Z	0·5	2	0·9	1·2	
	26	A	10,000	5,000	5,000	1,000	600
		Z	1·4	0·9	0·7	0·17	0·05
	37	A	3,000	30	3,000	1,000	
		Z	0·2	0·02	0·05	0·07	
5 Aug.	10	A	2,000	100	5,000	3,000	2,000
		Z	0·1	0·05	0·6	0·4	0·2
	18	A	500	3,000	3,000	300	
		Z	0·05	0·17	0·20	0·12	

TABLE 4. INDIVIDUAL RESULTS : hydrogen burner (night)

Date (1955)	Distance (km)	A = AgI nuclei/l Z = ZnS particles/l			
3 Aug.	10	A	4,000	2,000	2,000
		Z	1·5	1·6	2·0
	35	A	1,000	200	300
		Z	0·4	0·3	0·5
	47	A	200	300	300
		Z	0·19	0·2	0·10
5 Aug.	10	A	2,000	200	
		Z	0·1	0·02	

TABLE 5. TEMPERATURE SOUNDINGS

Date (1955)	A = altitude (m) T = temp. (°C)							
5 July	A	0	300	1,100	1,300	1,800		
	T	11	9	3	6	3		
12 July	A	0	1,600	1,800	2,000			
	T	11	− 1	4	3			
17 July	A	0	1,500	1,600	1,800			
	T	12	1	4	3			
18 July	A	0	200	500	600	1,300	1,500	1,800
	T	9	8	6	5·5	0	4	2
20 July	A	0	900	1,250	1,800	2,400		
	T	10	1	0	4	1		
24 July	A	0	300	2,000				
	T	8	8	− 3				
25 July	A	0	200	500	1,300	1,600		
	T	7	9	8	2	1		
26 July	A	0	100	600	1,600	1,700		
	T	8	10	8	1	1		
3 Aug.	A	0	1,200					
	T	7	− 2					
5 Aug.	A	0	200	400	1,300	1,700		
	T	5	7	6	− 1	− 1		

The flux of zinc-sulphide particles was measured on five occasions on which the smoke plume was adequately explored. The average value was $3{\cdot}4 \times 10^{13}$ particles/hr, with a standard deviation of 9×10^{12} particles/hr. These values were computed from the zinc-sulphide record, the aircraft flight path and the wind speed.

(*a*) *Decay*

The measurements of flux of freezing nuclei from the kerosene burner, after various exposure times by day, are displayed in Fig. 1 as crosses. The points are somewhat scattered but in general indicate a flux of about 10^{14} nuclei per gram AgI after a few minutes' exposure, the flux decreasing by about an order of magnitude per hour. Also plotted in Fig. 1, are the results of the previous measurements (Smith *et al.* 1955) shown as circles. The average ambient temperature in the 1953 experiments was + 14°C whereas in the 1955 experiments it was + 3°C. According to Bolton and Qureshi this difference of 11°C should have resulted in the rate of decay in the 1955 trials being less than half that observed in the 1953 trials. The scatter in the points of Fig. 1 prevents any accurate assessment of the rate of decay but no significant differences between the two sets of points can be seen, indicating that the ambient temperature had little if any effect on the rate of decay.

The measurements of flux of nuclei from the kerosene burner at night are shown on Fig. 1 as squares. They are closely grouped about a horizontal line, showing that there was no decay in the ice-nucleating properties by night. The cause of the decay by day is thus shown to be the daylight.

The measurements at night exhibit less scatter than those made in the daytime. Similarly, the individual measurements of each series also exhibit less scatter by night than by day, three-fourths of the individual measurements being within a factor of 2 of the geometric mean of the series by night, and within a factor of four by day. The reduced scatter at night suggests that the inherent accuracy of the measurements is reasonably good and the measured variations in decay in the daytime may be real. However, attempts to correlate the rate of decay with meteorological parameters such as cloud cover and humidity have been unsuccessful.

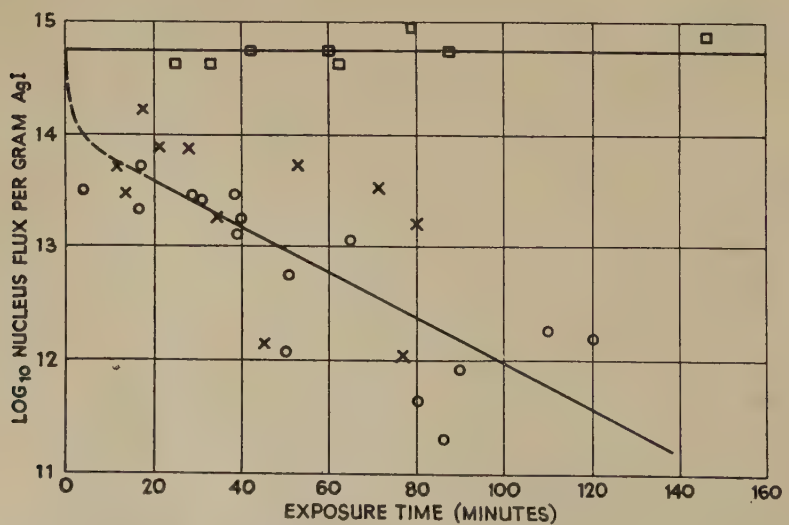

Figure 1. Variation of flux of freezing nuclei with time of exposure in the free atmosphere. Kerosene burner : × present trials – day; □ present trials – night; ○ 1953 trials – day.

The measurements made at night are close to 5×10^{14} nuclei per gram AgI, for exposure times between 24 min and 146 min. If this figure is taken to be the effective nucleus output of the burner, it indicates that, in the daytime, decay in the first few minutes' exposure might be at a greater rate than that subsequently prevailing.

The night-time measurements of the flux of freezing nuclei from the hydrogen silver-iodide burner are displayed as squares on Fig. 2, the daytime measurements of the previous trials (Smith *et al.* 1955) being shown as circles. Fewer measurements were made than with the kerosene burner, and the number of measurements per series was also less. Consequently the results do not reveal whether these nuclei experienced no loss or a very slow loss of properties at night. However, they clearly show that the rate of decay was at least very much less than in the previous daytime experiments.

(*b*) *Trajectory*

The width of the smoke plume, at varying distances from the generators, is given in Table 1, normal values being 5-km width at 12-km distance and 10-km width at 40-km distance. These widths are about half those previously published (Braham *et al.* 1952, Smith and Heffernan 1954, Crozier and Seely 1955) for smoke released from the ground on flat plains.

The altitude of the top of the plume was found on six occasions to be between certain limits, as shown in Table 1. These limits are displayed on Fig. 3 as crosses joined by vertical lines. On one occasion, 5 July, the top of the plume at 34-km distance was above 1,800 m. This is shown as a cross and an arrow on Fig. 3. At distances not greater than

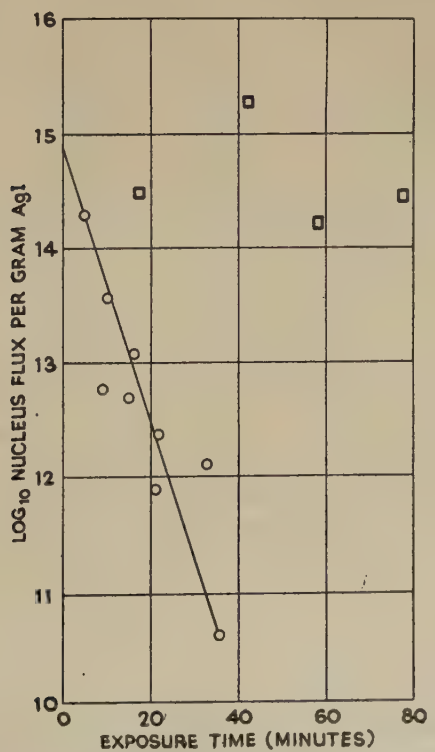

Figure 2. Variation of flux of freezing nuclei with time of exposure in the free atmosphere.
Hydrogen burner : □ present trials – night; ○ 1953 trials – day.

10 km, zinc sulphide was detected only on one occasion out of four at an altitude exceeding that of the generator; this was on 25 July, in a light wind at night. At distances of 31 to 34 km the height of the top of the smoke was, on two occasions, about 1,500 m. On one occasion, when it was above 1,800 m, there was a temperature inversion at the height of the smoke generators, and zinc sulphide was detected both above and below it. The bottom of the smoke plume fell below 500-m altitude within 10 km of the generators. On the other flights the aircraft usually flew through the smoke at an altitude of 1,000 m.

The measurements of the height limits of the top of the smoke plume in the 1953 trials are shown for comparison in Fig. 3 as pairs of circles separated by vertical lines, or by circles and arrows on two occasions when the altitude of the top was not established. While there was considerable variation from day to day, in general it appears that the smoke rose to about 1,500 m at a distance of 10 km from the generators, and thereafter rose very slowly if at all. The conditions in which these 1953 measurements were made are summarized in Table 6. Some of these data have been given by Smith *et al.* (1955) and some by Crozier and Seely (1955).

TABLE 6. 1953 TRIALS : HEIGHT OF TOP OF PLUME

Date	Distance from smoke generators (km)	Height of top of plume (m)	Lapse rate (°C/1000 m)	Inversion height (m)	Cloud	Temperature (°C at 250 m)
9 July	8	600- 750	9	> 1,200	7/8Sc	9
23 July	2	150- 200	9	> 1,200	6/8Sc	12
28 July	16	1,500-1,700	10	1,800	Fc	15
29 July	6	1,300-1,500	9	2,100	2/8Ac	10
30 July	11	1,300-1,500	10	> 1,800	Nil	16
11 Aug.	7	750- 900	9	> 900	5/8Sc	8
23 Oct.	23	900-1,100	9	1,400	6/8CiSt	15
23 Oct.	45	900-1,200	9	1,400	6/8CiSt	17
6 Nov.	56	▷ 1,100	9	2,400	5/8Sc	14
10 Nov.	8	1,500-1,800	10	1,500	Nil	14
10 Nov.	34	1,500-1,800	10	1,500	Nil	15
12 Nov.	19	> 2,700	10	3,000	1/8Ac 5/8Ci	24

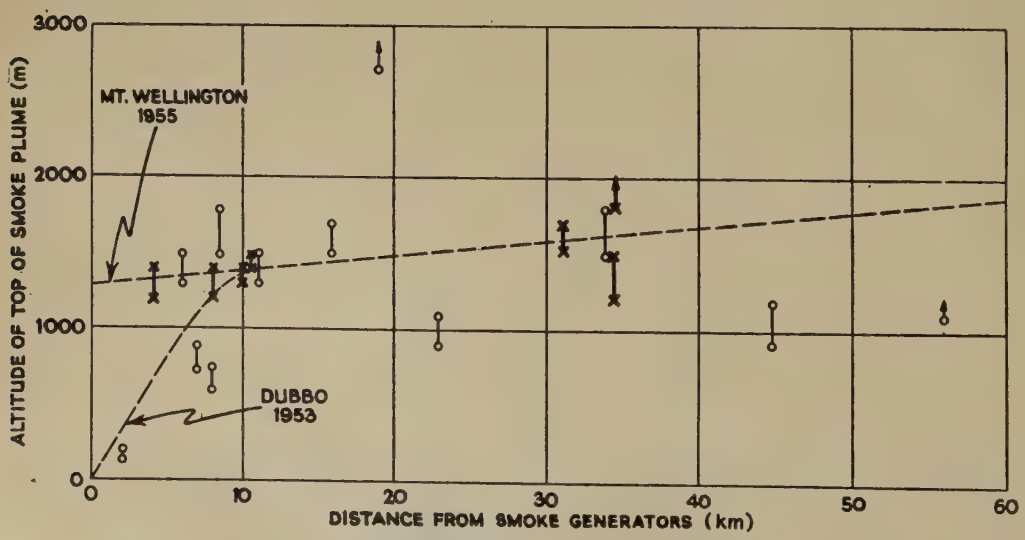

Figure 3. Altitude of top of smoke plume.

present trials, 1953 trials.

From Fig. 3 it can be seen that the smoke released from a 1,300-m mountain top did not extend higher, at distances exceeding a few kilometres, than that released in previous trials from flat plains. However, the height of the top of the plume no doubt reflects the local conditions, which in this case usually involved temperature inversions.

6. Conclusions

The measurements reported above show that the decay of the ice-nucleating properties of silver-iodide smoke is due to daylight, while any effect of ambient temperature is too small to be detected by this method. It is to be hoped that this knowledge will stimulate the development of freezing nuclei which are not affected by daylight.

If attempts to produce an adequate concentration of freezing nuclei at cloud level are limited by the daytime decay of silver iodide, then in some circumstances it may be preferable to operate the smoke generators at night. However, due regard must be paid to the local conditions, since night-time cloud seeding may present other problems such as altered incidence of suitable clouds or limitations of smoke trajectory, e.g., by ground inversions.

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551.521.61

The influence of the 15 μ carbon-dioxide band on the atmospheric infra-red cooling rate

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Summary

The upward and downward radiation flux and cooling rate are calculated for the 15 μ band of carbon dioxide. Results are obtained for three different carbon-dioxide concentrations from the surface of the earth to 75 km, and for six frequency intervals covering the band. The infra-red absorption measurements of Cloud (1952) are used for calculations, on a high-speed electronic computer, by a method which takes account of the pressure and Doppler broadening, the overlapping of the spectral lines, and the variation of the intensity and half-width of the spectral lines with temperature and pressure. The numerical integration is performed over intervals that are never larger than 1 km and average values over layers are not used. The cooling rate for the present atmospheric carbon-dioxide concentration is greater than 1°C/day from 24 km to 70 km and is greater than 4°C/day from 38 km to 55 km. The sum of the ozone and carbon-dioxide cooling rates is greater than 4°C/day from 33 km to 57 km and agrees reasonably well with the heating due to ozone absorption. The results for different carbon-dioxide concentrations indicate that the average temperature at the surface of the earth would rise by 3·6°C if the carbon-dioxide concentration were doubled and would fall by 3·8°C if the carbon-dioxide concentration were halved, on the assumption that nothing else changed to affect the radiation balance.

1. Introduction

The results of calculations of the infra-red radiation flux and the heating and cooling rates in the atmosphere for the spectral region from 12 μ to 18 μ are presented in this paper. The magnitude of the radiation flux in this region is determined largely by the properties of the 15 μ carbon-dioxide band. For atmospheric layers very close to the surface of the earth a correction must be introduced for absorption by water vapour. The radiation flux has been computed for three different carbon-dioxide concentrations when there are clear skies, and when clouds are present at one or the other of two different heights. All calculations were made from the surface of the earth to a height of 75 km. The assumed variation of temperature and pressure with height agrees closely with the values given by the Rocket Panel (1952).

The calculations are based on the carbon-dioxide absorption measurements of Cloud (1952). The radiation flux was obtained from these laboratory measurements by the same method as was used by Plass (1956a) for the spectral region near the 9·6 μ ozone band. Complete details of the method are given in that paper. However, for the carbon-dioxide calculations it was necessary to take into account the variation with temperature of the absorption in a given frequency interval; this was calculated from the partition function and the known dependence on temperature of the intensity and half-width of the spectral lines. The atmosphere is not divided into homogeneous layers for the purpose of this calculation, as this procedure may introduce large errors in the determination of the cooling rates. At each stage a careful estimate is made of the accuracy of the calculation.

An excellent survey of previous work on the carbon-dioxide absorption problem is given by Elsasser and King (1953). The pioneer work of Ladenburg and Reiche (1911) has been extended in recent years by Callendar (1941), Elsasser (1942), Kaplan (1950), and Yamamoto (1952). Kaplan (1952) and Elsasser and King (1953) have made the most extensive previous calculations of the absorption of the carbon-dioxide band.

* This paper was completed while the author was on leave of absence at Michigan State University. At present with Systems Research Corporation, Van Nuys, California.

2. Method for calculation of atmospheric transmission from laboratory data

The extensive and detailed study of the 15 μ carbon-dioxide band by Cloud (1952) was made on the large 100-foot absorption cell built by Professor John Strong at Johns Hopkins University. Path lengths up to 600 ft could be obtained by means of a six-fold optical path. The measurements were made with a wide slit so that the individual spectral lines were not resolved and the mean transmission could easily be obtained for various intervals. All the measurements made by Cloud are shown in Fig. 1 for the entire spectral interval from 12 μ to 18 μ. The logarithm of the fractional absorption, A, is plotted against the logarithm of the product of the pressure, p, and path length w, used in the laboratory. All the measured points lie on a single universal absorption curve within the experimental accuracy of about two per cent.

However, the entire interval from 12 μ to 18 μ must be divided into smaller intervals in order to calculate the radiation flux in the atmosphere. The black-body intensity is replaced by an average value over the interval in question in order to perform the integration over frequency. Since the black-body intensity varies with frequency very slowly compared with the extremely rapid variations of the absorption coefficient, it can be shown numerically that the substitution of an appropriate average value for the black-body intensity introduces no appreciable error for frequency intervals as large as one micron. For this reason the interval from 12 μ to 18 μ was divided into six intervals, each one micron wide, and all quantities were calculated for each of these intervals. The absorption measurements of Cloud (1952) for each of these intervals are shown in Figs. 2, 3 and 4.

The atmospheric radiation flux and cooling rate were calculated from these measured absorption curves by the method described by Plass (1952a). Details of the procedure, a discussion of the influence of the assumed temperature distribution on the final results, and detailed calculations of the probable error of the final results are described by Plass (1956a) for the similar calculations for ozone. In particular, the following corrections were made to the transmission function : (1) for Doppler broadening above 50 km (Plass and Fivel 1953); (2) extrapolation of experimental results to very small path lengths (Plass 1952a, 1954; Plass and Fivel 1955b); (3) multiplication of the path length by a numerical factor varying with the transmission, in order to obtain the diffuse radiation (Plass 1952b); (4) for the variation of line strength and half-width with temperature (the procedure for making this correction is given in the Appendix).

It was assumed that the mixing ratio for carbon dioxide is a constant independent of height up to 75 km. This is a good approximation for the average carbon-dioxide amount, except possibly very near the ground. The fluctuations of the carbon-dioxide mixing ratio are small compared to those of water vapour and ozone and should have very little influence on the radiation flux. The pressure and density curves given by the Rocket Panel (1952) were used with the necessary corrections for the small differences between their temperature and our assumed temperature.

The following temperature distribution was assumed for all the carbon-dioxide calculations :

$$
\begin{aligned}
T &= 288 - 6z & 0 &\leqslant z \leqslant 13 \text{ km} \\
T &= 210 & 13 &\leqslant z \leqslant 22 \text{ km} \\
T &= 210 + 3(z - 22) & 22 &\leqslant z \leqslant 43 \text{ km} \\
T &= 273 & 43 &\leqslant z \leqslant 54 \text{ km} \\
T &= 273 - 3(z - 54) & 54 &\leqslant z \leqslant 75 \text{ km},
\end{aligned}
$$

where z is the height in kilometres. These values agree closely with those given by the Rocket Panel (1952) and should represent an average temperature distribution in middle

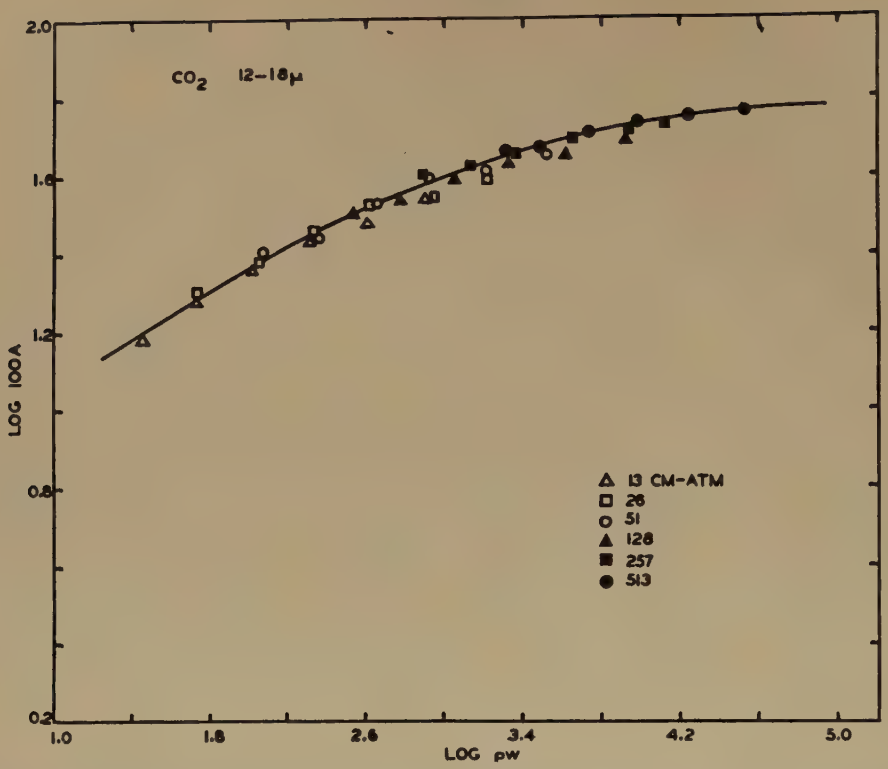

Figure 1. The fractional absorption, A, of CO_2 as a function of the product of the pressure, p, and the optical path length used in the laboratory, w, for the spectral interval from 12 μ-18 μ. The measurements were made by Cloud (1952).

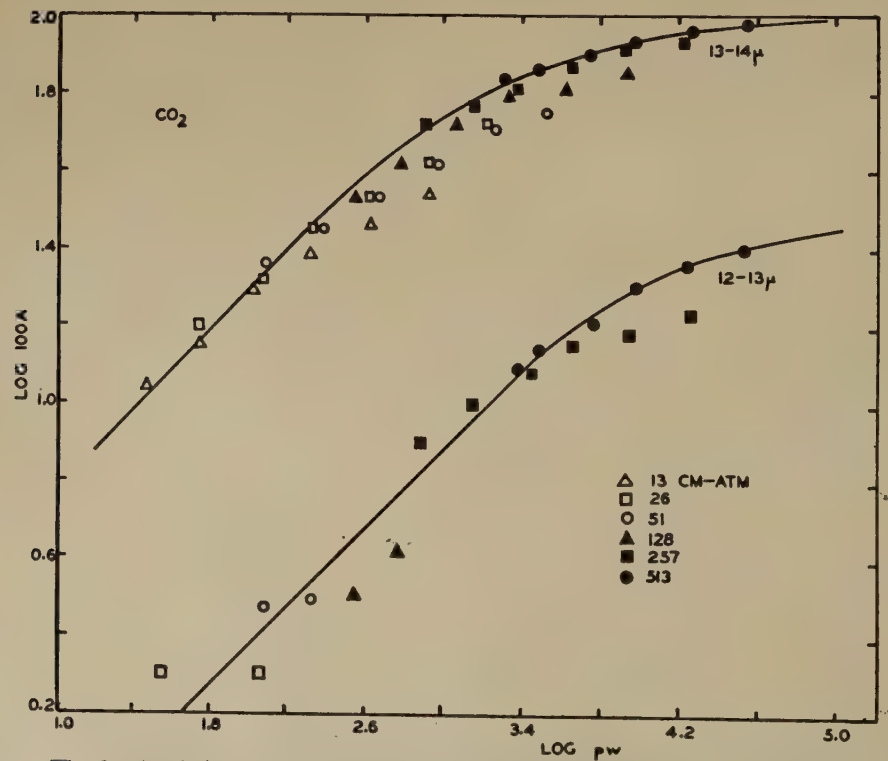

Figure 2. The fractional absorption of CO_2 for the spectral intervals from 12 μ-13 μ and from 13 μ-14 μ as measured by Cloud (1952).

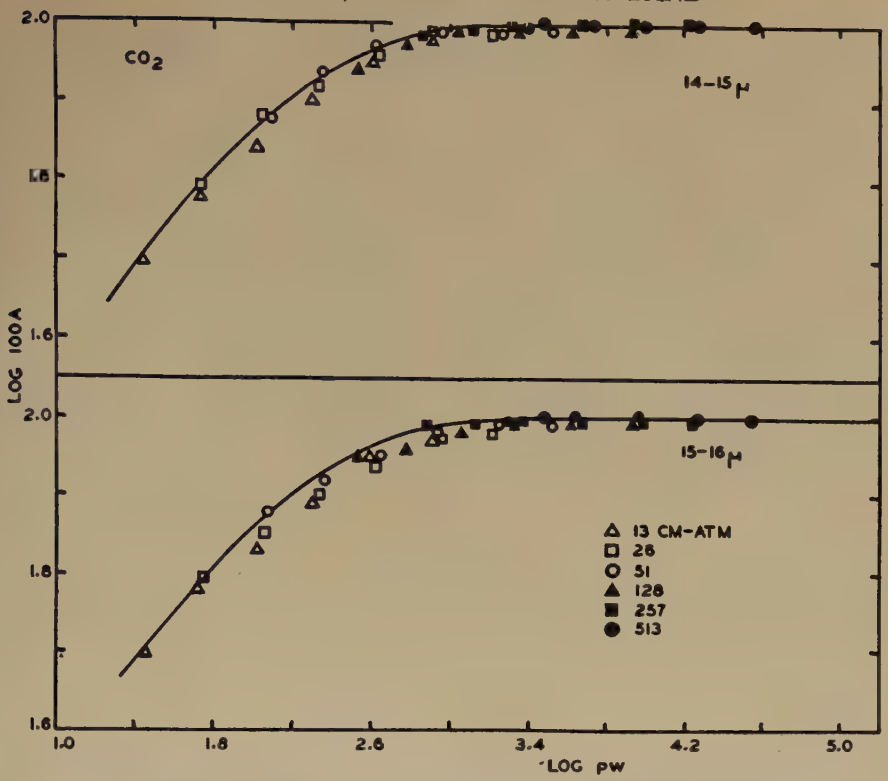

Figure 3. The fractional absorption of CO_2 for the spectral intervals from 14 μ-15 μ and from 15 μ-16 μ as measured by Cloud (1952).

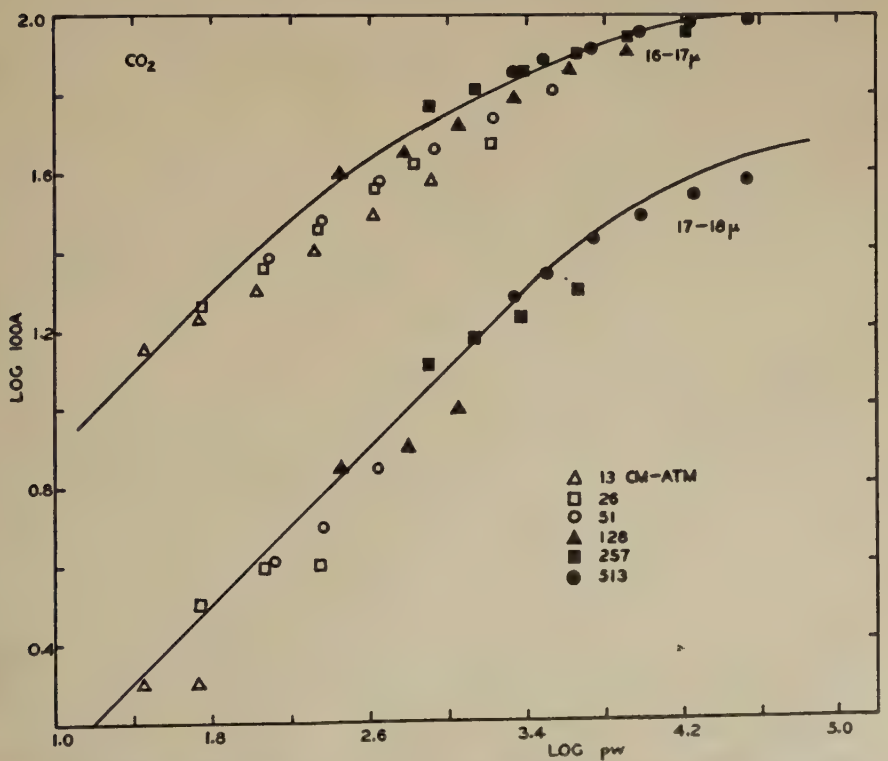

Figure 4. The fractional absorption of CO_2 for the spectral intervals from 16 μ-17 μ and from 17 μ-18 μ as measured by Cloud (1952).

latitudes. A more accurate result can be obtained from the same number of steps in the numerical calculation if the atmosphere can be divided into regions each of which has a linear variation of temperature with height, as here.

The necessary integrations over the transmission functions (Eqs. 1-3, Plass 1956a) were performed numerically for values of the integrand at intervals of 1 km. When the integrand was a rapidly varying function, even smaller intervals were chosen. The atmosphere was not divided into homogeneous layers for the purpose of this calculation, as this procedure may introduce large errors in the cooling rate. For each of the six spectral intervals, and for each of the three different carbon-dioxide concentrations, it was necessary to calculate the transmission function for approximately 3,000 pairs of values of z_0 and z_1 between the surface of the earth and 75 km, where z_0 and z_1 are the heights of the lower and upper level respectively. The entire numerical calculation, including the corrections to the transmission function, was coded for the MIDAC high-speed digital computer at the University of Michigan.

3. The infra-red radiation flux and cooling rate

The results of the calculation of the infra-red radiation flux and cooling rate by the above method are shown in Figs. 5-8. The upward and downward radiation flux is given for the entire spectral interval from 12 μ-18 μ and for three smaller intervals. In order to present as many of the results as possible in a reasonable space, the results for three pairs of intervals have been combined. Thus instead of showing the flux for the 12 μ-13 μ and 17 μ-18 μ intervals separately, as calculated, only the total for both intervals is given; the variation with height of the results for these two intervals is almost the same and no essential information is lost by combining the results in this manner. For the same reason the results for the 13 μ-14 μ and 16 μ-17 μ intervals were combined, as were those for the 14 μ-15 μ and 15 μ-16 μ intervals.

Figs. 5 and 6 give the results for the present concentration of CO_2 ($c = 5 \times 10^{-4}$; 0·033 per cent by volume) when the temperature correction in the Appendix is made (solid curve) and is not made (dashed curve).

The change in the upward flux with height depends on the transmission of the spectral region in question. If the transmission were unity in a given spectral interval, the higher atmospheric layers at lower temperatures would be unable to change the upward flux and it would be constant with height. If the transmission were zero, the upward flux would be equal to the black-body intensity at every height and, for our assumed temperature distribution, would have a minimum from 13 km to 22 km and maxima at the ground and from 43 km to 54 km.

The upward flux for various spectral intervals as shown in Fig. 5 is close to these limits. For the 12 μ-13 μ; 17 μ-18 μ interval the transmission is near unity; the upward flux is nearly constant with a value above 13 km only 8 per cent lower than the value at the ground. For the 13 μ-14 μ; 16 μ-17 μ interval the transmission has an intermediate value at the lower altitudes; between 13 km and 22 km the upward flux falls to almost one-half of its value at the ground. It increases slightly above 22 km, but this rise is very small, since the transmission is so near unity for paths at the higher altitudes that higher temperatures there can have little effect. For the 14 μ-16 μ interval the transmission is nearly zero up to 15 km; here the upward flux is very close to the black-body curve. Above this height the upward flux increases as the atmospheric temperature increases to a value 28 per cent above its minimum value. The upward flux for the entire range from 12 μ-18 μ exhibits the characteristics of a curve for a transmission function with a value intermediate between zero and unity.

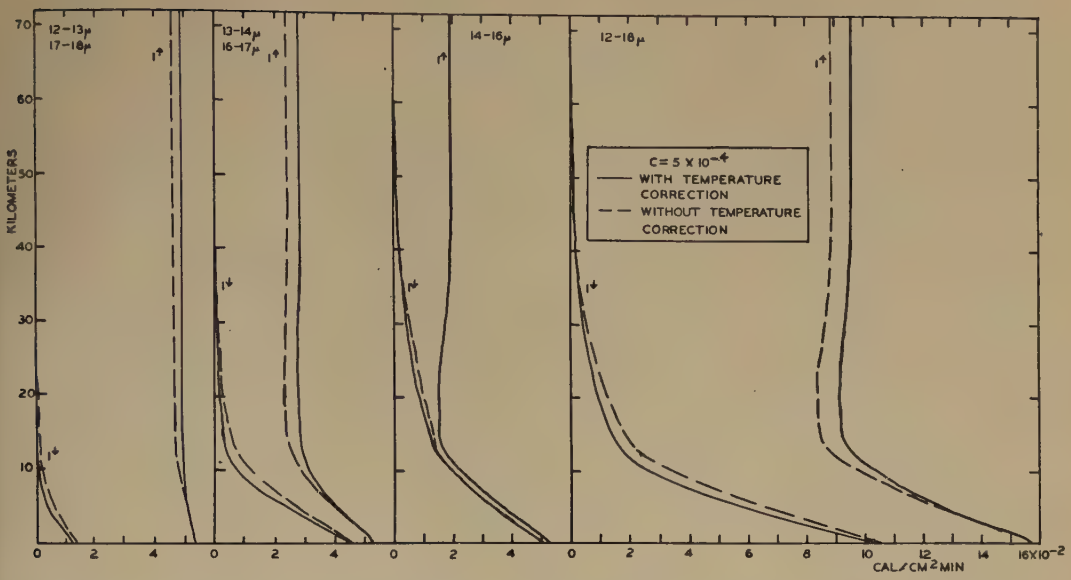

Figure 5. The upward and downward radiation flux as a function of height for the combined frequency intervals from 12 μ-13 μ and 17 μ-18 μ; from 13 μ-14 μ and 16 μ-17 μ; from 14 μ-16 μ; and for the entire interval from 12 μ-18 μ. The temperature correction described in the Appendix was made in the calculations for the solid curves and was omitted for the dashed curves. The present CO_2 concentration ($c = 5 \times 10^{-4}$; 0·033 per cent by volume) was assumed for this calculation. The difference between the two curves for the upward flux for the 14 μ-16 μ interval cannot be shown on the scale of this graph.

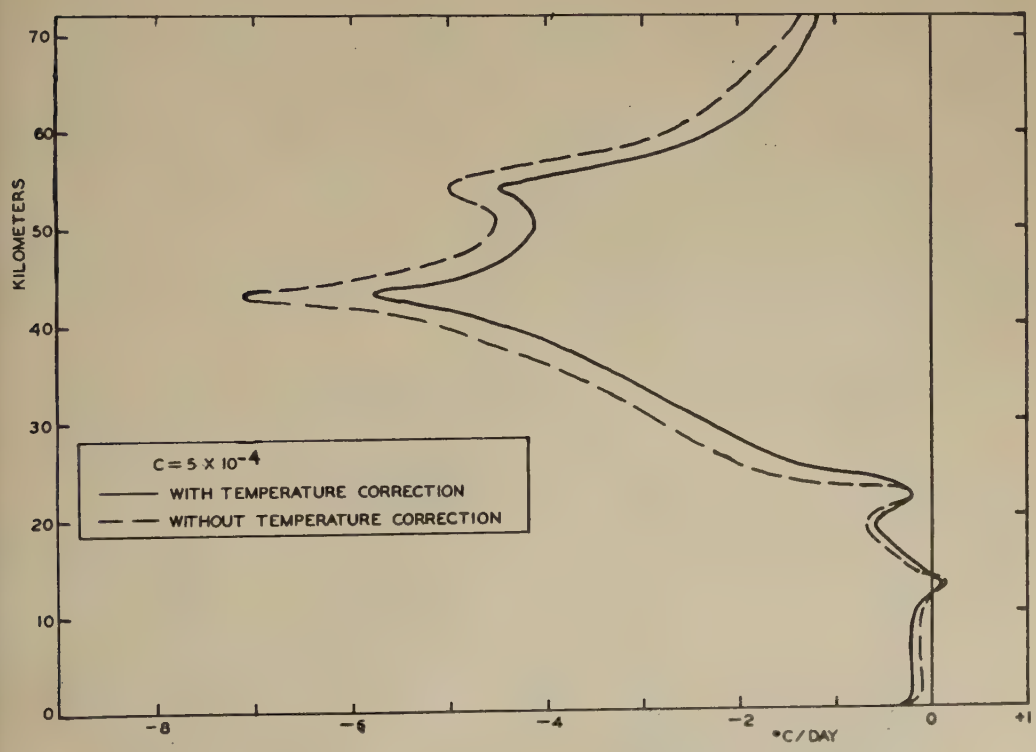

Figure 6. The cooling rate in °C/day for the entire frequency interval from 12 μ-18 μ.

There would be no downward flux if the transmission function were unity in a given frequency interval, since the atmosphere could not radiate at these frequencies. If the transmission function were zero, the downward flux would equal the black-body intensity for the appropriate temperature at a given height. For the interval from 12 μ-13 μ; 17 μ-18 μ the downward flux is quite small and has an appreciable value only below 10 km. For the interval from 13 μ-14 μ; 16 μ-17 μ the downward flux has a considerably larger value. For the interval from 14 μ-16 μ, the downward flux is larger than for either of the previous intervals and approaches the black-body flux below 13 km. The upward and downward flux would have an identical value equal to the black-body intensity below 13 km for the 14 μ-16 μ interval, if the transmission function were actually zero for all possible path lengths below 13 km. Actually for sufficiently small path lengths, even near the surface of the earth, the transmission function approaches unity; this is the reason for the small difference between the upward and downward flux in this case.

The cooling rate for the entire CO_2 band from 12 μ-18 μ is shown in Fig. 6. From the ground up to 10 km the cooling rate is of the order of a few tenths of a degree Celsius per day. This rate rises rapidly from 22 km to 43 km and reaches a maximum of 5·8°C per day at 43 km. The cooling rate is above 4·0°C per day from 38 km to 55 km. The rate decreases rapidly at higher altitudes. The maxima and minima in this curve at 13 km, 22 km, 43 km and 54 km arise from the discontinuities in the derivative of the assumed temperature curve at these altitudes (Plass 1956a). Such discontinuities seldom, if ever, actually exist in the atmosphere. Therefore the values in the neighbourhood of these maxima and minima should be smoothed out slightly when comparisons are made with the actual atmosphere. As discussed above, this temperature distribution was assumed, as it greatly increases the accuracy of numerical calculations carried to a given number of figures; it approximates closely the actual temperature distribution.

A comparison of the cooling rates of the ozone and carbon-dioxide bands is given Table 1. The ozone cooling rates are those calculated by Plass (1956a) for the ozone-concentration curve *L* and the temperature curve *B* as defined in that article. This temperature curve is very close to the one used for the present carbon-dioxide calculations, except that it does not have discontinuous derivatives at certain heights. A small correction has been made to the carbon-dioxide results between 40 km and 60 km to bring them more closely into agreement with the results that would have been obtained had the temperature curve *B* been used. The largest relative contribution from the ozone is at 45 km where it contributes 30 per cent to the total energy radiated from both ozone and carbon dioxide. The carbon dioxide is much more important than the ozone in cooling the atmosphere above 55 km.

TABLE 1. COOLING RATES IN THE STRATOSPHERE

Height (km)	Ozone (°C/day)	Carbon Dioxide (°C/day)	Total (°C/day)
25	0·1	1·3	1·4
30	0·5	2·4	2·9
35	1·3	3·3	4·6
40	1·8	4·5	6·3
45	2·1	4·9	7·0
50	1·5	4·6	6·1
55	0·8	4·0	4·8
60	0·3	2·3	2·6
65	0·1	1·7	1·8
70	0·0	1·3	1·3

The absorption of solar energy by ozone at these altitudes has recently been calculated by Johnson (1953) and Pressman (1955). Since the heating rates vary greatly with latitude and season (Pressman 1955), it is difficult to make quantitative comparisons. However, it is clear that the heating and cooling rates agree qualitatively and have their maxima at the same altitude (45 km) with approximately the same value of 7·0°C/day. The cooling due to H_2O is not included in Table 1. Differences between the heating and cooling rates could be ascribed to the possible additional cooling effect of H_2O and to uncertainties in the calculations themselves.

The correction for the variation of the line intensities and half-widths with temperature, discussed in the Appendix, greatly complicates the calculation and introduces an additional factor of uncertainty in the final results. In order to study the effect of this correction, the calculations were made with and without the temperature correction, for each of the three CO_2 concentrations. The results are shown in Figs. 5 and 6 for the present CO_2 concentration ($c = 5 \times 10^{-4}$). The principal temperature correction is due to the increase in line intensity of the stronger lines of the band with increasing temperature over the temperature range that occurs in the atmosphere. Since the laboratory measurements of the transmission function are made at room temperature, the temperature-corrected transmission function is greater than the uncorrected value for the colder regions of the atmosphere. Thus the results are changed most when the temperature correction is made at altitudes from 10 to 30 km where the atmospheric temperature is relatively low.

The value of the upward flux at high altitudes is 5 per cent higher for the interval from 12 μ-13 μ; 17 μ-18 μ, and 14 per cent higher for the interval from 13 μ-14 μ; 16 μ-17 μ, when the temperature correction is made than when it is not made. The upward flux has a larger value since the transmission function is closer to unity after the temperature correction. Thus the upward flux cannot follow so well the dip in the black-body curve in the lower stratosphere. The change for the 14 μ-16 μ interval cannot be shown on the scale of Fig. 5, since the transmission function is virtually zero in the lower part of the atmosphere in both cases.

For the downward flux the difference between the calculated values with and without the temperature correction is largest in each case near the base of the stratosphere. In the intervals from 12 μ-13 μ; 17 μ-18 μ, and from 13 μ-14 μ; 16 μ-17 μ, most of the downward radiation reaching, for example, 8 km, originates in the colder region of the atmosphere from 10 km to 30 km. It is not surprising therefore that the increased values for the transmission after the temperature correction is made give values for the downward flux that are as little as half those obtained without the correction. This is the only region of the atmosphere where the temperature correction must be made in order to obtain even approximately correct values of the flux. At the surface of the earth so much of the downward radiation comes from the lower, warmer layers of the atmosphere that the difference between the two curves is less than 10 per cent for the interval from 12 μ-13 μ; 17 μ-18 μ, and is very small for the other intervals.

The cooling curve with and without the temperature correction is given in Fig. 6. Above 14 km the cooling rates are slightly smaller when the temperature correction is made. However, the shapes of the two curves are remarkably similar and the percentage deviation of the two curves is not large except below 14 km where in any case the cooling rate due to CO_2 is not of great importance. Thus it would appear that reasonably accurate calculations of the cooling rate above 14 km can be made if desired without introducing the complexities of the temperature correction.

The infra-red flux and cooling rates are shown in Figs. 7 and 8 for three different CO_2 concentrations, 0·025, 0·05 and 0·10 per cent by weight or 0·0165, 0·033 and 0·066 per cent by volume. The lowest and highest values correspond to halving or doubling

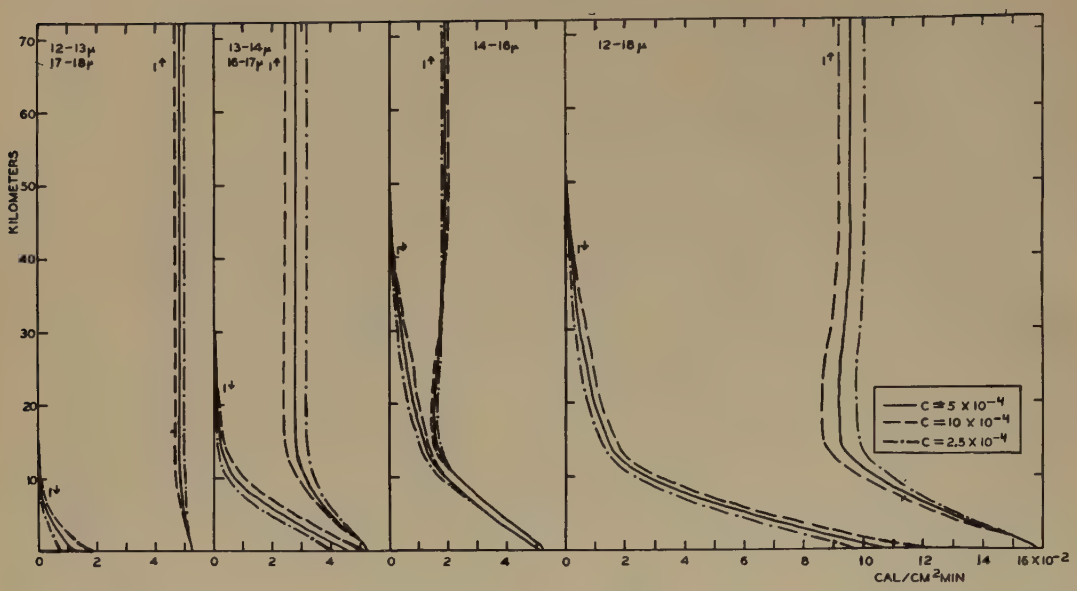

Figure 7. The upward and downward radiation flux as a function of height for the combined frequency intervals from 12 μ-13 μ and 17 μ-18 μ; from 13 μ-14 μ and 16 μ-17 μ; from 14 μ-16 μ; and for the entire interval from 12 μ-18 μ. Curves are given for the following CO_2 concentrations: $c = 5 \times 10^{-4}$ (0·033 per cent by volume); $c = 10 \times 10^{-4}$ (0·066 per cent by volume); $c = 2{\cdot}5 \times 10^{-4}$ (0·0165 per cent by volume). The temperature correction described in the Appendix was made in the calculation of all these curves.

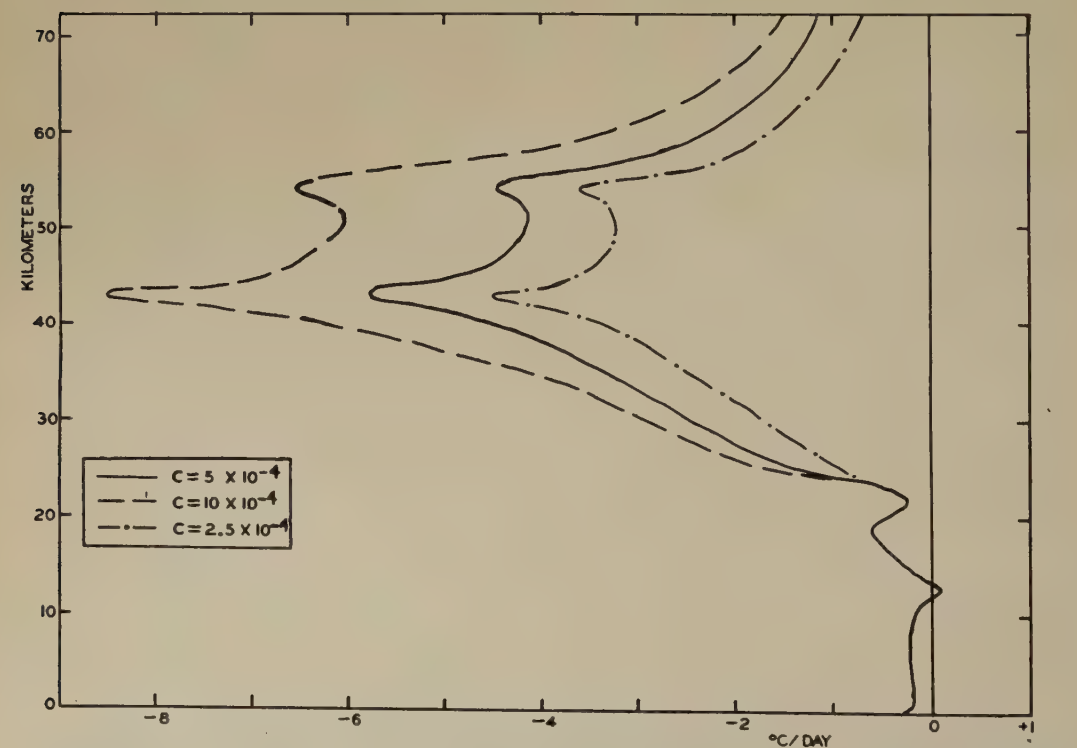

Figure 8. The cooling rate in °C/day for the entire frequency interval from 12 μ-18 μ. Curves are given for the following CO_2 concentrations : $c = 5 \times 10^{-4}$ (0·033 per cent by volume); $c = 10 \times 10^{-4}$ (0·066 per cent by volume); $c = 2{\cdot}5 \times 10^{-4}$ (0·0165 per cent by volume). The temperature correction described in the Appendix was made in the calculation of all these curves. There is no difference in the cooling rates below 24 km within the accuracy of the calculation and only one curve is shown in this region.

the present atmospheric CO_2 amount. The temperature correction has been made to all the results given in these figures, since the complete procedure had been coded for the MIDAC electronic computer.

For the intervals from 12 μ-13 μ; 17 μ-18 μ, and from 13 μ-14 μ; 16 μ-17 μ, the upward flux is smaller the larger the CO_2 concentration, since the smaller transmission function allows the flux to approach closer to the black-body flux at the temperature minimum from 13 km to 22 km. The same is true for the interval from 14 μ-16 μ below 32 km. Above this height the upward flux is slightly greater for larger CO_2 concentrations, since the transmission function is still sufficiently small in this region for the upward radiation to increase slightly owing to the higher temperature in the upper part of the stratosphere. Similar remarks apply to the downward flux. In this case the larger CO_2 concentrations give the larger values for the downward flux.

The cooling rates are given in Fig. 8 for these three different CO_2 concentrations. There was no definite difference within the accuracy of the calculation, in the cooling rates below 24 km, and only one curve is shown in this region. At higher altitudes the cooling rate increases appreciably if the CO_2 concentration changes. At the maximum of the cooling curve at 43 km, the cooling rate increases from 5·8°C/day to 8·5°C/day if the CO_2 concentration is doubled.

Numerous checks were made during the progress of these calculations in order to determine the accuracy of the final results. The procedure was similar to the one used by Plass (1956a) for ozone. The upward and downward flux is determined with a percentage error smaller than that of the original laboratory absorption measurements. Cloud's estimate of a 2 per cent error in the laboratory measurements of the transmission function may be somewhat optimistic. A conservative estimate of the error in the upward and downward flux, taking into account the various corrections made in the course of the calculations, is 3 per cent. However, the relative changes of the curves for different concentrations are determined with considerably greater accuracy than this. The probable error of the cooling rate is estimated by introducing arbitrary variations into the original transmission functions and calculating their influence on the final result. The probable error obtained in this manner is about 10 per cent below 20 km, increasing to 30 per cent at 50 km and becoming rather uncertain above 60 km. Again the relative differences between the various curves should be considerably more accurate than their magnitude.

4. Temperature variations at the surface of the earth

The change in the equilibrium temperature at the surface of the earth with CO_2 concentration can be estimated from the corresponding variations of the radiation flux, if it is assumed that nothing else changes to alter the radiation balance when the CO_2 amount varies. In order to obtain the temperature change it is also assumed that an additional amount of radiant energy equal to 0·0033 Cal cm^{-2} min^{-1} would be radiated to space, if the average temperature of the earth's surface were to increase by 1°C. This number cannot be calculated accurately until a detailed study has been made of the H_2O spectrum, but it represents the best value that can be given with our present knowledge of this spectrum. When a more accurate value for this number is obtained in the future, all the temperature changes given here should be multiplied by the ratio of the old to the new value.

When the CO_2 amount is doubled and halved the change in the downward flux at the surface of the earth is 0·0119 and 0·0125 Cal cm^{-2} sec^{-1} respectively. Thus, in order to restore equilibrium the surface temperature must rise by 3·6°C if the CO_2 concentration is doubled and the surface temperature must fall by 3·8°C if the CO_2

concentration is halved. It is interesting to note that the four intervals from 12 μ-13 μ, 13 μ-14 μ, 16 μ-17 μ and 17 μ-18 μ contribute about equally to this temperature change. The interval from 14 μ-16 μ does not contribute since it is opaque near the ground at all these concentrations. The argument has sometimes been advanced that the CO_2 cannot cause a temperature change at the surface of the earth because the CO_2 band is always black at any reasonable concentration. This argument is true for the lines near the centre of the band from 14 μ-16 μ, but neglects completely the important contribution of the lines farther from the band centre.

This same method shows that the average temperature is lowered 2·2°C and 1·3°C at the upper surface of a cloud at 4 km and 9 km respectively, if the amount of CO_2 in the atmosphere is halved. This temperature change at the upper surface of a cloud could be the cause of the increased precipitation at the beginning of a glacial period (Plass 1956b).

If the CO_2 concentration is doubled, Arrhenius (1896) calculated a temperature increase of 6°C, Hulbert (1931) of 4°C, and Callendar (1938) of 2°C. The value of 3·6°C obtained from the present study is based on more accurate experimental data than were available previously and it was possible to perform the elaborate calculations only because an electronic computer was available. Such factors as the weak lines that are relatively far from the band centre and the change in intensity and half-width of the spectral lines with temperature and pressure were taken into account in the present calculations.

The radiation calculations predict a definite temperature change for every variation in CO_2 amount in the atmosphere. These temperature changes are sufficiently large to have an appreciable influence on the climate. Plass (1956b) has considered the various factors that control the CO_2 content of the atmosphere and oceans and has discussed the current status of the carbon dioxide theory of climatic change.

Acknowledgments

This research was supported by the Office of Naval Research, Washington. I wish to thank Mr. L. Leopold for his extensive and careful work in calculating the radiative flux in the early stages of this study. His work made it possible to code the final calculation for the MIDAC high-speed electronic computer.

Appendix

Variation of transmission function with temperature

Laboratory measurements of the absorption of atmospheric gases should ideally be made over the range of temperatures that occur in the atmosphere. This would eliminate the need to make a theoretical estimate of the variation of the transmission function with temperature. However, to our knowledge no suitable measurements have yet been made at temperatures below room temperature.

First let us assume that the temperature decreases linearly with height between z_0 and z_1. Without any loss of generality z_0 may be set equal to zero. Then the temperature dependence is given by

$$T = T_0 - a z \qquad (1)$$

and the pressure dependence by

$$p = p_0 (T/T_0)^{Mg/aR} \qquad (2)$$

The temperature dependence of the line intensity, S, and, to a smaller extent, the temperature variation of the half-width of the spectral line, α, cause the transmission function to vary with temperature. The theoretical variation with temperature of both these factors is known. From the Maxwell-Boltzmann distribution law and the rotational partition function, it follows that the temperature dependence of S is given by the expression

$$\frac{S(T)}{S(T_0)} = \frac{T_0}{T} \exp\left[-\frac{E_J}{kT} + \frac{E_J}{kT_0}\right] \quad . \quad . \quad . \quad (3)$$

where T_0 is a standard temperature, $S(T)$ and $S(T_0)$ are the line intensities at the temperatures T and T_0 respectively, E_J is the energy of the initial state of the transition, and k is the Boltzmann constant. Other terms in the general expression for S which change very slowly with T are omitted from Eq. (3), since numerical integration shows that their effect is very small for the range of temperatures in the atmosphere.

From kinetic theory it follows that

$$\alpha = \alpha_0 (p/p_0) (T_0/T)^{\frac{1}{2}} \quad . \quad . \quad . \quad . \quad (4)$$

where α_0 is the half-width at a standard pressure and temperature, p_0 and T_0. Although the dependence of α on pressure has been verified, a square-root temperature dependence should not be expected for all lines. However, the variation of α with temperature introduces only a very small correction.

For a single spectral line, the result of the usual method for the calculation of the absorption in the atmosphere integrated over frequency, A, (Plass 1954, Plass and Fivel 1955a) is that

$$A(u_0, u_1) = \left\{4 \sec\theta \int_{u_0}^{u_1} S\,\alpha\, du\right\}^{\frac{1}{2}} \quad . \quad . \quad . \quad (5)$$

for the strong-line approximation and that

$$A(u_0, u_1) = \sec\theta \int_{u_0}^{u_1} S\, du \quad . \quad . \quad . \quad . \quad (6)$$

for the weak-line approximation, where

$$u = \int_z^\infty c\,\rho\, dz,$$

$$A(u_0, u_1) = \int_0^\infty [1 - \tau(u_0, u_1)]\, d\nu,$$

θ is the angle the radiation makes with the vertical, c is the ratio of the density of the radiating gas to the total density, ρ, and τ is the transmission function. Eqs. (5) and (6) are valid regardless of the functional dependence of S on height. The results for overlapping spectral lines are discussed later in this section.

In order to obtain the integrated absorption, A, for the strong-line approximation, substitute Eqs. (1), (2), (3) and (4) into Eq. (5). Define an effective temperature, T_e, between z_0 and z_1 so that the integrated absorption, $A(z_0, z_1)$, for a radiating gas in a hypothetical atmosphere with the same temperature, T_e, at all heights, is the same as for the actual radiating gas with variable temperature. The effective temperature, T_e, is a function of both z_0 and z_1. It can be calculated by setting the expression for A, obtained

from Eq. (5) when the entire atmospheric layer is at the temperature T_e, equal to the expression for A, derived above, which is valid when the temperature decreases linearly with height.

In this manner it is found that T_e can be determined from the other known constants of the problem, from the equation

$$\int_1^{\beta} \exp\left(-\frac{E_J y}{kT_0}\right) y^{-\frac{2Mg}{aR}+\frac{1}{2}}\, dy = \left(\frac{T_0}{T_e}\right)^{\frac{3}{2}} \frac{aR}{2Mg}\left[1-\beta^{-\frac{2Mg}{aR}}\right] \exp\left(-\frac{E_J}{kT_e}\right). \quad . \quad (7)$$

where $$\beta^{-1} = 1 - (az_1/T_0).$$

The value of the effective temperature, T_e, can be obtained as a function of z_0 and z_1 from the numerical solution of Eq. (7). Some typical results are shown in Fig. 9, where T_e is plotted against the temperature at the boundary of the layer. The upper set of three curves gives T_e for a layer between z_0, where the temperature is 293°K, and z_1, where the temperature has the value given by the abscissa. The lower set of three curves gives T_e for a layer extending from z_0, where the temperature has the value given by the abscissa, to z_1, where the temperature is 218°K. Each set of curves is drawn for three different values of the ratio $m = E_J/kT_{293}$. Since the value of m does not change by more than a factor of three when averaged over a frequency interval of 0·1 μ or more in the range from 12 μ-18 μ, Fig. 9 shows that the value of T_e is insensitive to the assumed value of m.

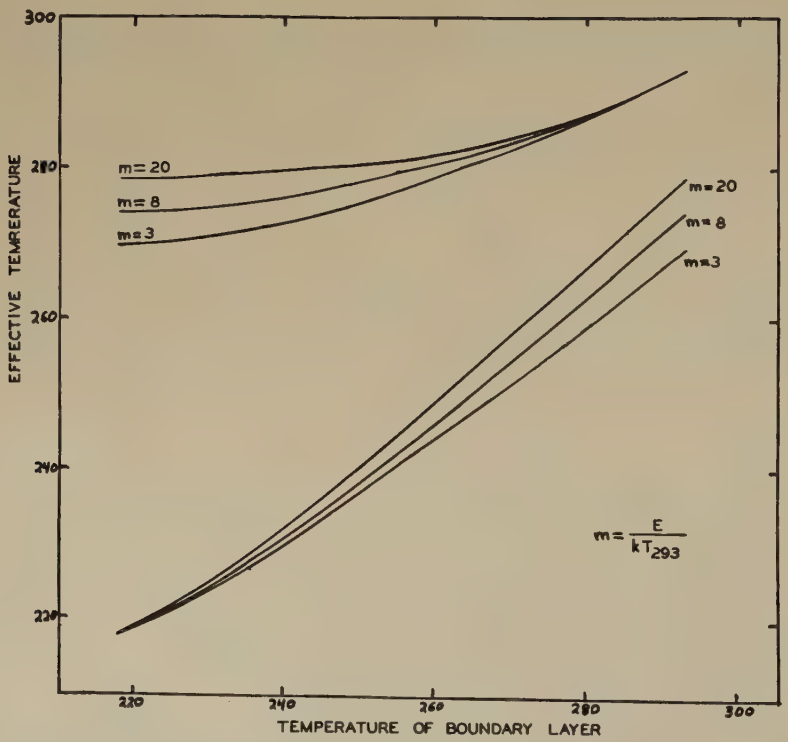

Figure 9. The upper set of three curves gives the value of the effective temperature, T_e, for an atmospheric layer extending from a fixed height where the temperature is 293°K to a height where the temperature has the value given by the abscissa. The lower set of three curves gives the value of T_e for an atmospheric layer extending from a height where the temperature has the value given by the abscissa to a fixed height where the temperature is 218°K. Each set of curves is drawn for three different values of the ratio $m = E_J/kT_{293}$, where E_J is the energy of the initial state of the transition.

When the equation for the weak-line approximation, Eq. (6), is used instead of Eq. (5), a similar derivation to the one given above shows that

$$\int_1^{\beta} \exp\left(-\frac{E_J y}{kT_0}\right) y^{-\frac{Mg}{aR}-1}\, dy = \left(\frac{T_0}{T_e}\right)\frac{aR}{Mg}\left[1-\beta^{-\frac{Mg}{aR}}\right]\exp\left(-\frac{E_J}{kT_e}\right) \quad . \quad . \quad (8)$$

The results for T_e are very similar to those obtained from Eq. (7).

Eqs. (7) and (8) have been derived for a single spectral line. However, the identical equations are also valid for the Elsasser model of a band in the limits of the strong- and weak-line approximations. This result can be derived from the equations for the Elsasser model given by Plass and Fivel (1955b). Since T_e does not depend on the degree of overlapping of the spectral lines for the Elsasser model in the limits of the strong- and weak-line approximations, it appears that T_e cannot vary appreciably with the degree of overlapping of the spectral lines for any model of a band.

Equations similar to Eqs. (7) and (8) were derived for the cases when the temperature increases linearly with height or when there are several regions with a constant temperature in one region and a linear increase or decrease in another. In practice, for the temperature distribution assumed for this calculation, it is never necessary to use more than two regions with different lapse rates for the calculation of T_e. If more than two regions exist with different lapse rates between z_0 and z_1, the regions above the first two make a negligible contribution.

The values of the ratio E_J/kT, where T is the temperature at which the laboratory absorption measurements were made, were calculated from a tabulation of the carbon dioxide lines kindly made available to us by Dr. L. D. Kaplan in advance of publication. The region from 12 μ to 18 μ was divided into intervals 0·1 μ wide. In each of these 60 intervals the average energy of the lines was computed. It was weighted either with the square-root of the line intensity or the first power of the intensity depending on whether the absorption for the entire interval followed the strong-line or the weak-line approximation respectively for the path length between z_0 and z_1. The average energy for the six intervals of one micron width from 12 μ to 18 μ was then calculated from the averages for the 0·1 μ intervals. The deviations of the values for the 0·1 μ intervals from the average for the 1 μ intervals were relatively small. For example, for the interval from 12 μ to 13 μ the average deviation of the results for the smaller intervals was only 10 per cent of the average value for the entire interval.

The calculated value of E_J/kT was between 3 and 8 for each of the six intervals of 1 μ width. Fig. 9 shows that T_e is insensitive to the value of E_J/kT. Thus it is possible to obtain the value of T_e for any combination of values of z_0 and z_1 with an estimated probable error of no more than a few degrees Kelvin.

Curves similar to those in Fig. 9 were plotted at intervals of 1 km from the surface of the earth to 75 km. They were calculated from Eqs. (7) and (8) or, when the temperature did not decrease linearly with height, from equivalent equations. The effective temperature, T_e, was determined in this manner for each pair of values of z_0 and z_1. The temperature correction to the integrated absorption was then calculated from Eqs. (3), (4) and the known equations for the dependence of the absorption on these variables (Plass 1952a, 1954; Plass and Fivel 1955b). Fortunately it was possible to code a programme of calculations for the MIDAC high-speed digital computer to calculate this correction entirely automatically for each pair of z_0 and z_1.

All the carbon-dioxide calculations were made both with and without this temperature correction. The correction was largest for the intervals, from 12 μ-13 μ and 17 μ-18 μ, that are far from the centre of the band, and was smallest for the region from 14 μ-16 μ at the centre of the band.

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Short-period variations in the atmospheric electric potential gradient

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Summary

Measurements made with 2 field mills, 100 m apart, along the direction of the surface wind, have been used to obtain information about the charges responsible for variations in the potential gradient over periods of the order of minutes. In general these variations are caused by the horizontal motion of wind-borne space charge contained within the first few hundred metres of atmosphere. In fair weather, there are effects probably caused by locomotive steam, and others perhaps to be associated with convection cells. In overcast weather there is evidence for concentrations of negative charge in the base of the thicker portions of the cloud. In showers, the main result is that the pattern of the potential gradient is very much affected by space charges from point discharge.

1. Introduction

The vertical potential gradient in the atmosphere close to the earth's surface has been measured in a number of different ways at very many places. The diurnal and annual variations and the very rapid variations caused by lightning flashes have all been extensively studied. However, much less attention has been given to variations over periods of the order of minutes (the main work being that of Simpson (1949), concerning ' field patterns ') and it is with such variations that we shall be concerned in the present work.

Kasemir (1950) has shown that the long-period variations of potential gradient can be described most simply in terms of the current flowing. Such a description, however, is not appropriate for variations whose periods are not long compared with the relaxation time of the atmosphere, which has a value at the surface of 6 to 40 min, depending on the local conductivity. It is then necessary to describe the phenomena in terms of electric charges and electrostatic forces which can be pictured as lines of force having one end on a unit charge on the earth's surface and the other on a unit charge of opposite sign somewhere in the atmosphere. The magnitude of the potential gradient is then proportional to the density of these lines of force. Changes in the potential gradient must, therefore, be associated with a movement of the space charges.

Measurements of the potential gradient at any number of points on the earth's surface cannot give, uniquely, the distribution of the charges responsible. However, simultaneous measurements at two places not far apart can give very useful information. For example, if the variations of potential gradient at the two places show no similarity, then the charges responsible for the variations must be very local. Alternatively, if there are similar variations, then the time delay between them must represent the velocity of the charge resolved along the line joining the points. If this line is along the direction of the surface wind we can determine whether the responsible charge is carried by the wind and, if the wind velocity varies with height, we may be able to obtain some idea as to the height of the wind-borne charge.

2. Apparatus

For the measurement of potential gradient two identical field mills were constructed; these have been described separately (Mapleson and Whitlock 1955). The output from each mill was connected to a micro-ammeter for visual observation and to a critically damped galvanometer of period 0·2 sec for photographic recording. The minimum change of potential gradient detectable was about 0·5 v/m.

In order to avoid disturbances from buildings and from atmospheric pollution, the investigation was carried out at the Durham University Observatory. Eight sites were chosen on a circle of approximately 50 m radius, centred on the Observatory building. The two mills were set up at the most suitable pair of sites so that the line joining the mills was approximately along the surface wind and their distance apart about 100 m. In the discussion of the records the two mills are referred to as the ' upwind ' and ' down-wind ' mills. If the charges moved with the surface-wind speed then the time intervals for corresponding effects would be some seconds. 960 m distant in a direction of 096° from the Observatory are the University Science Laboratories where additional simultaneous measurements were made, when required, with an agrimeter (Chalmers 1953).

For use with the mills a ' sky photometer ' was constructed, this consisted of a vertical tube with a photo-cell at the bottom which gave an output depending on the brightness of the overhead sky.

The wind velocity during each recording was obtained from the daily records of the Observatory Dines pressure tube anemometer, which was specially recalibrated at the end of the work. The anemometer is 18·5 m above ground level (the Meteorological Office considers its effective height to be 10 m) and the wind velocity as measured by this anemometer will be termed the ' surface-wind velocity ' and denoted by V_s (direction in degrees and speed in m/sec).

When the potential gradient is sufficiently large, point discharge can occur at trees and other points, including in this case the anemometer mast. In order to measure the point-discharge current, a point was set up at a height of 19·0 m above the ground and the current through it measured by a galvanometer. Both the anemometer mast and the point were directly above the Observatory building and so approximately midway between the two mills.

3. Interpretation of results

In most of the records there was found a close similarity between the potential gradients at the two mills when placed along the line of the surface wind. The time intervals between the corresponding maxima and minima and zero values showed good consistency, the average values being of the same order as calculated for the surface wind. An example of such a record reproduced in Fig. 1 shows an average interval of 33 sec corresponding to a charge speed of 3·1 m/sec whilst the surface wind speed was 2·5 m/sec. This recording was made during a period of light drizzle.

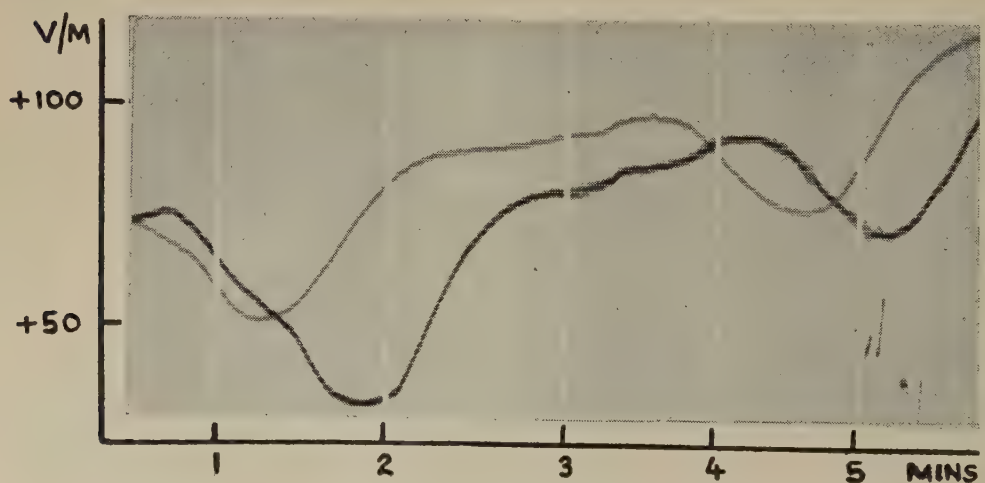

Figure 1. Record showing potential-gradient variations at two sites 100 m apart.

Fig. 2 is a 'scatter diagram' of the time intervals during a period of steady snow and much variation of potential gradient. Inserted in the diagram are also the time intervals to be expected if the charges concerned moved with the mean surface wind speed or with the calculated geostrophic wind speed (discussed later).

For all the records for which time intervals could be measured a charge speed was obtained, referred to as V_c, assuming the direction of charge motion to be the same as that of the surface wind. The ratio of the average value of V_c to the surface wind velocity V_s, is referred to as R. For all the records from which R could be obtained, its value lay between 0·7 and 2·0 with an average value of about 1·25. If the form of the variation of wind speed with height were known, it would be possible to use the value of R to obtain the height of the charges concerned. Unfortunately, upper-air measurements were not made at or near Durham thus the values of R can give only rough estimates of height.

On some occasions, when the wind was in the right direction, it was possible to compare the time interval between a mill at the Observatory and the agrimeter at the Laboratories, with the time interval between the two mills. The ratio of the intervals was found to be close to the ratio of the distances, confirming the interpretation of the intervals as the charge speed. Fig. 3 illustrates this in a case which is discussed later.

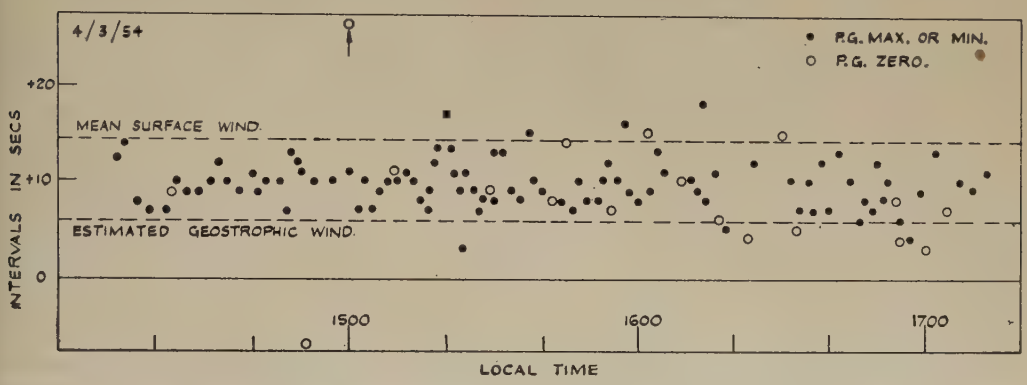

Figure 2. Time intervals between corresponding potential-gradient variations at two sites 100 m apart.

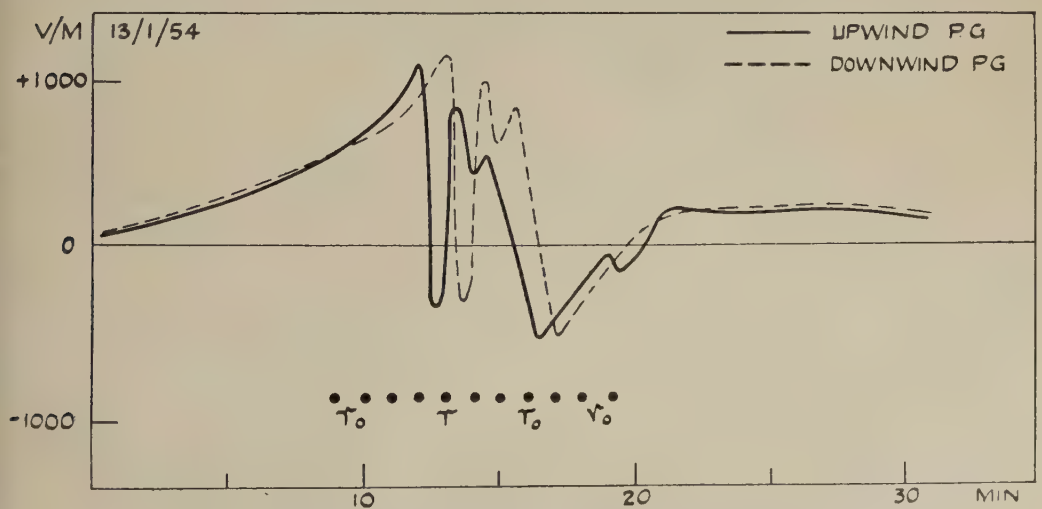

Figure 3. Simultaneous potential-gradient variations at two sites 960 m apart.

The diagrams show the potential gradients as measured, the state of the sky as observed, the surface wind as measured, the charge speed as deduced from the time intervals, the ratio R, and indications of precipitation. In certain cases the point discharge and the sky photometer traces are also given. In some cases the values of V_c and R are given twice, the second in brackets; the bracketted value of V_c refers to the charge speed resolved along the direction of cloud motion instead of along the direction of the surface wind. The bracketted value of R is the ratio of this value of V_c to V_s.

4. Fine and fair weather

Recordings were made on 7 occasions of truly fine weather with cloudless skies and to these may be added 3 occasions with a small amount of cirrus cloud. In all cases the potential gradient was remarkable steady, showing that any charge existing in the lower atmosphere was uniformly distributed in any horizontal cross-section. Mean values of the potential gradient were obtained for each minute and the average change, irrespective of sign, was obtained from one minute to the next. This average change, expressed as a percentage of the mean potential gradient, gives a measure of the ' field unrest ' and is referred to as U. For about 10 hr of fine-weather recording the average value of U was about 2·9 per cent. With such small variations in the potential gradient the time intervals for the two mills were difficult to measure. For such cases in which measurements could be made the average value of the ratio R was 1·13, which indicates an average height of about 50 m for the charges responsible for the variations.

In fair weather of the type which gives rise to non-raining cumulus or altocumulus cloud, showing the presence of convection, recordings were made on 10 occasions with a total time of about 22 hr. The average value of U was 12·8 per cent, much greater than in fine weather. In general it was found that the value of U increased with the wind speed, in agreement with some comparable results of Israël (1943). The average value of the ratio R was 1·18, corresponding to a height of about 60 m. No correlation was found between the potential-gradient changes and the cumulus clouds above, showing, as does the value of R, that the variations were not caused by charges in these clouds.

It appears likely that the extent of potential-gradient variation and hence the value of the ' field unrest ' depends on the amount of convection in the lower atmosphere.

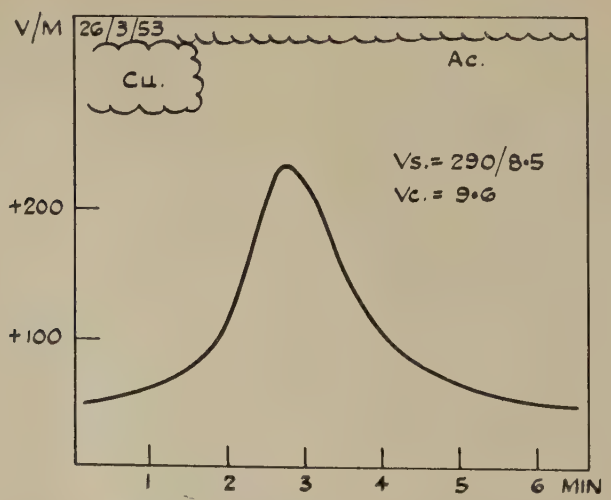

Figure 4. A fair weather ' field pulse.'

It was not possible to test this hypothesis with any degree of certainty since the nearest places to Durham at which upper-air temperature measurements were made are some hundreds of kilometres away, but such results do indicate that the greatest values of U were obtained in conditions of super-adiabatic lapse rates with vertical instability and much turbulence.

Frequently in fair weather, but not in truly fine weather, a temporary increase of the potential gradient to about twice its normal value was observed, with a return to normal in 2 or 3 min. An example is given in Fig. 4. Time intervals showed that the charges responsible were moving at speeds slightly greater than the surface-wind speed and visual observation showed no connection with clouds. The sky photometer only rarely showed any effect. These ' field pulses,' as they will be called, gave rapid changes in potential gradient and so must be caused by quite local concentrations of charge. Such concentrations could not survive long in a turbulent atmosphere, so their origin must have been fairly local. An examination of the direction of the wind when the pulses occurred suggested that they were caused by the positive charge emitted with locomotive steam, as found by Kelvin (1860), Israël (1950) and others, coming from a railway line about 1·5 km distant. If it is assumed that the charge responsible consists of an infinite line charge at right angles to its direction of motion, the shape and magnitude of the field pulse makes it possible to calculate the height and density of the charge. Average values are 160 m and $1{\cdot}5 \times 10^{-6}$ C/metre length of line charge. In Fig. 4 the calculated height is 115 m and the charge is $8{\cdot}6 \times 10^{-7}$ C/metre. The height obtained by these calculations is greater than would be expected from the time intervals but the assumptions made are only approximate. It can be seen that lateral dispersion of a line would tend to give a greater width to the pulse and hence a greater height by calculation than actually occurred.

Another feature of the fair-weather measurements, again absent in truly fine weather, was the frequent occurrence of ' cusp ' variations such as is shown in Fig. 5, the pattern elements being repeated a few times, dying out and reappearing later. In a great many cases a definite periodicity was noticed, varying from 1·5 to 9·0 min with an average value of 6·2 min. The time intervals between the two mills were measured and gave an average value of R of 1·18. From the periodicity and the charge velocity it was possible to obtain the distances between the charge elements responsible and these varied from 1,150 m to 6,100 m with an average value of 3,400 m. Although these cusp variations occurred only when cumulus clouds were present, no connection could be observed between the cusps and the clouds. It is suggested that the cusps are connected with convection ' cells ' or ' bubbles.' Giblett *et al.* (1932) found a cell structure 1-3 km wide and James (1952) found bubbles 0·1-2·5 km in diameter; others have given sizes

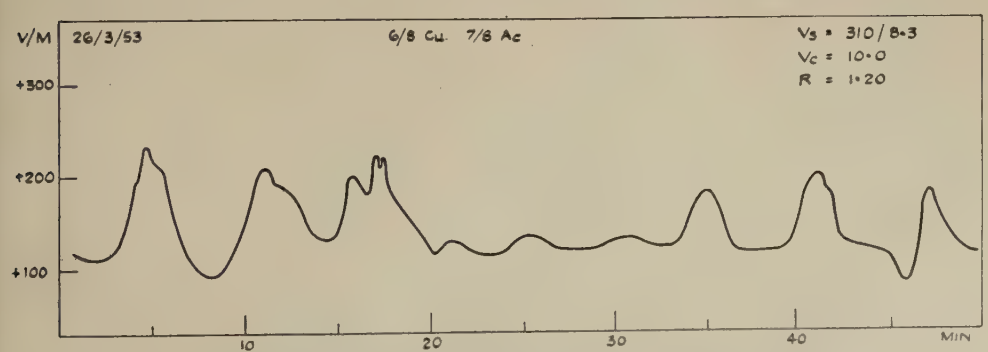

Figure 5. Periodic potential-gradient variations during fair weather.

of the same order. The cells or bubbles appear to have about the same degree of regularity as the cusp variations of potential gradient. If there is indeed a connection between the cusps and convection the mechanism might be that of a vertical column of positive charge carried up in the updraught of the convection process, the positive charge originating near the surface from the electrode effect. The potential-gradient changes of the cusps are of the right order of magnitude to agree with this picture. The correctness or otherwise of this interpretation of the cusp variations remains to be verified by a more detailed investigation involving, for example, measurement of the vertical component of the wind.

In fair weather, other than with cumulus clouds, little of interest was found except on three occasions when there was a negative potential gradient. These all occurred with the wind between ENE. and SE. and in conditions of high relative humidity. On these three occasions there probably was emission of negative ions from high-tension cables (Chalmers 1952), a process which appears responsible for negative potential gradients in mist and fog.

5. Mist and fog

In mist and fog, as Chalmers (1952) has shown, negative potential gradients often occur at Durham when the wind is in a direction from N. to SSW. via E. The effects may be caused by negative ions produced by insulation breakdown at high-tension cables. The records often show wavelike variations and a comparison with the photometer is shown in Fig. 6; in this case the potential gradient was positive. It is seen that the higher values of potential gradient are associated with a darker sky and so with a thicker mist. On other occasions when the potential gradient was negative a *darker* sky was associated with the highest *negative* values of the gradient. No satisfactory explanation of these results can be given. Values of R averaged 1·12, showing that the charges concerned were relatively low in the atmosphere.

6. Overcast sky

A number of measurements were made when the sky was overcast with stratus or stratocumulus cloud, both in cases without precipitation or with drizzle or light rain. Contrary to the opinions often expressed, the presence of such precipitation at the ground has very little effect on the potential gradient, for similar patterns occur with no precipitation and with drizzle. Fig. 7 gives an example with no precipitation and a positive potential gradient but in other cases, both with and without precipitation, there were periods of negative gradient. In Fig. 7 the sky-photometer output variations are also recorded and show good correlation with the potential-gradient variations. It should be noted that the highest potential gradient occurs with a bright sky, this being opposite

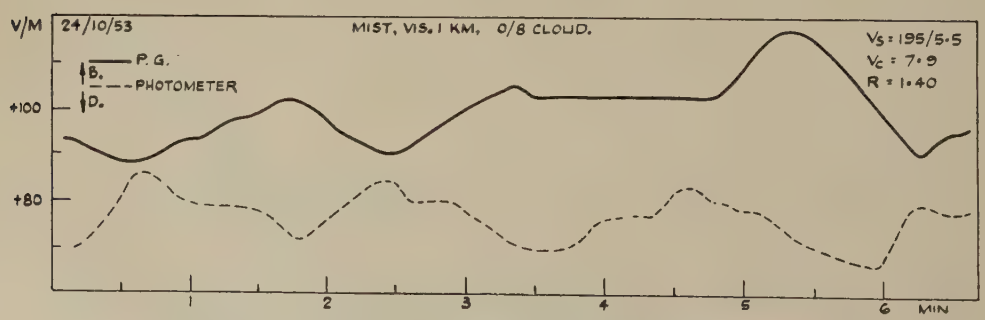

Figure 6. Variations in potential gradient and sky brightness in mist.

to the results for positive gradients in mist and fog. The time intervals associated with overcast sky variations were well marked and agreed in general with the hypothesis that the charges were in the cloud base.

It might be suggested that the correspondence of the potential gradient pattern with the cloud structure could be explained in terms of the difference in conductivity between cloud and air, and that the current flow would show a pattern beneath the cloud. Calculations show that the effects are larger than can reasonably be explained on this basis and negative potential gradients cannot be explained merely by differences in conductivity.

The results indicate some process of charge separation within the cloud, more effective in the thicker than in the thinner regions. Since the cloud base, and probably the whole cloud, was at a temperature above the freezing point, the process of charge separation cannot be the same as that which occurs in thunder-clouds. However, both processes give rise to negative charges on the lower part of the cloud. Two theories which have been put forward to account for thunderstorm electricity might be applicable in this case, namely that of Wilson (1929) and that developed by Gunn (1935) and on somewhat similar lines by Frenkel (1947). There does not appear to be sufficient evidence at present available to discuss usefully the application of either theory to the present results.

7. Continuous rain

In continuous rain from nimbostratus clouds, the potential gradients were often greater, and varied more rapidly, than with an overcast sky. There are examples in which there were corresponding changes in potential gradient at the Observatory and at the Laboratories, although the wind direction was far from the line joining the two. It appeared that the charges responsible were fairly high and might be located in the clouds.

8. Showers

Showers can be conveniently divided into three groups according to the values of the potential gradient reached, since these values determine whether point discharge occurred or not. When the potential gradient exceeded $\pm$ 400 v/m, point discharge occurred at the point set up at 19·0 m. When the potential gradient exceeded about $\pm$ 1,500 v/m point-discharge current appeared to flow from the anemometer mast, trees and the building, and possibly from trees and buildings further to windward. For a period the point was removed to distinguish its effect from those of the mast and the trees, etc.

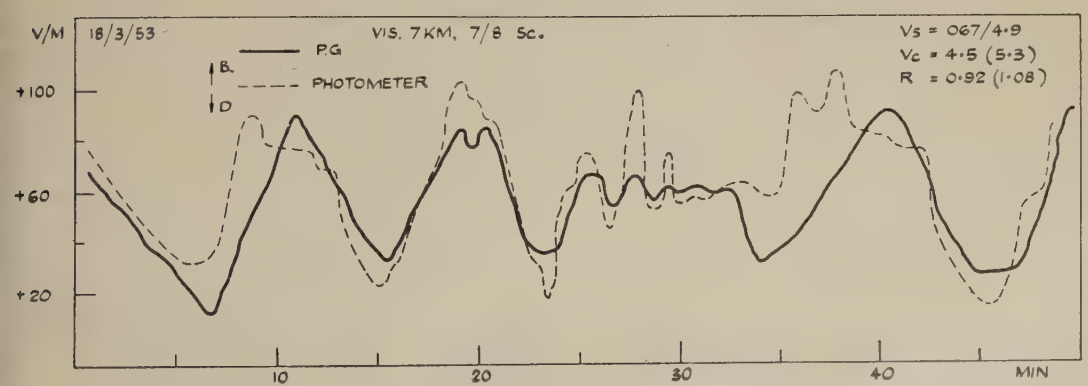

Figure 7. Variations in potential gradient and Sc cloud brightness.

For showers with low values of potential gradient the results generally show a period of negative gradient in the form of a V-pattern. An important feature of the records is that there are occasions of V-patterns with no precipitation reaching the ground. Alternatively, apparently quite ordinary shower clouds passed over and the potential gradient remained undisturbed at its normal positive value. An interesting case is shown in Fig. 8 where the very heavy rain coincided with the maximum rate of change of potential gradient. Whether precipitation does or does not fall appears not to be the determining factor in the separation of charge that gives rise to the patterns, though of course, there may be precipitation that does not reach the ground. It was also noticed that the V-patterns often showed an asymmetry, giving the impression that the negative charge responsible was trailing behind the cloud. The values of the ratio R, whilst greater than in fine weather, did not reach the values that might be expected if the charges responsible were entirely at cloud level.

When slight discharge occurred at the point this affected the downwind mill, but the effects at the upwind mill were in general similar to those previously described. Fig. 9 shows an example where the usual V-pattern has become a W-pattern, the central peak of the W coinciding with the heaviest rain, suggesting a positive charge on the rain. When the potential gradients became still greater the patterns became so complicated as to give little information. Some individual cases will be discussed in § 9 and the general effects of space charge in § 10.

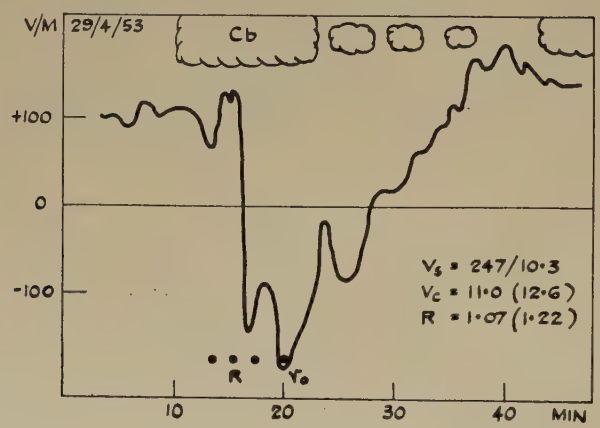

Figure 8. Potential-gradient variations beneath a small shower cloud.

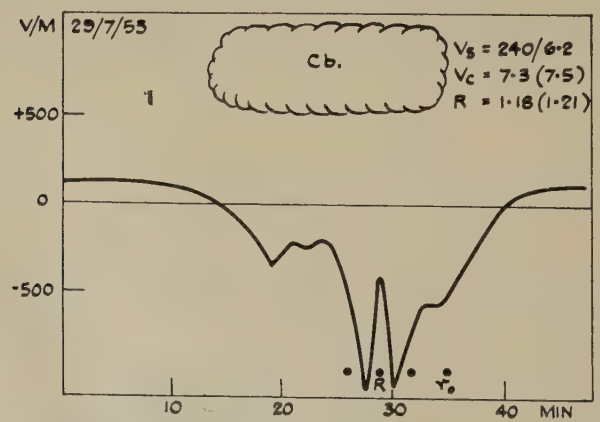

Figure 9. Potential-gradient variations beneath a shower cloud.

During thunderstorms the patterns were so complex that it is clear that the present methods are not suitable to give new information.

9. Special cases

In one remarkable case the potential gradients showed peak values some time after the cloud had passed over, leaving clear blue sky overhead. This suggested a considerable lag of the charge behind the cloud.

Fig. 3 is unusual in that the pattern must have been due mainly to positive charges. The cloud appeared to be a cumulonimbus in the final stages of dissipation, and it is this example which has been referred to in § 3 in connection with the persistence of a pattern over 960 m. The positive charge might well have been the remains of the upper positive charge of a typical shower cloud, the other charges having dispersed.

Another case shows a very rapid potential-gradient change, coinciding with the heaviest rain. If the rain carried a strong positive charge one might be able to account for the pattern.

On one occasion, during very high potential gradients associated with a sleet shower cloud, there occurred fifteen apparently instantaneous potential-gradient changes of about 10 v/m, which appeared similar to those usually associated with lightning flashes. However, no thunder was heard, nor was any reported nearer than 250 km. These small potential gradient changes may be similar to those reported by Gunn (1954) as occurring in thunderstorms.

10. Effect of point space charge

When point discharge occurs a charge is liberated opposite in sign to that of the potential gradient. This ' point space charge ' travels mainly with the wind and at places downwind of the discharging point it produces a reduction or even a reversal of the potential gradient. Since the wind near the earth's surface is usually gusty, particularly in the disturbed weather conditions in which point discharge occurs, the pattern of potential gradient produced may become ' wavy.' Fig. 10 shows the effect very clearly in a situation where there was only one point producing the point space charge. The record from the downwind mill shows a wavy pattern when, and only when, point discharge was occurring, whilst the upwind mill shows steady changes.

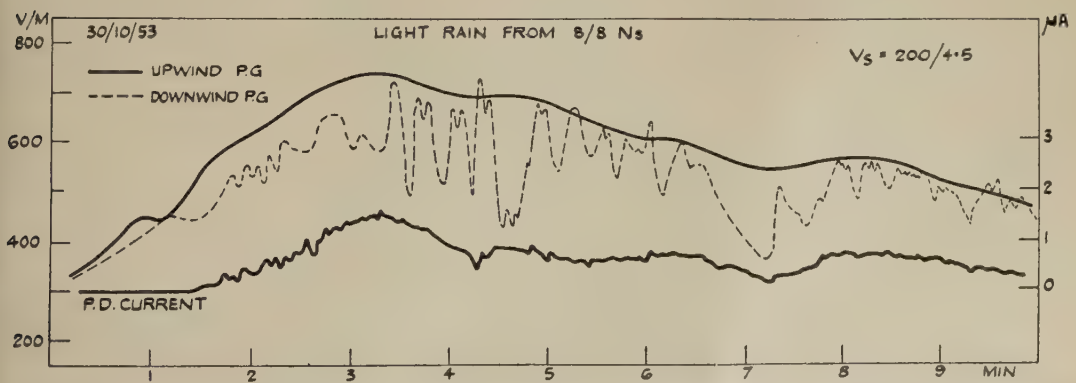

Figure 10. Potential-gradient variations 50 m upwind and 50 m downwind of a discharging point.

Fig. 11 shows the much more complex effects when the potential gradients were high enough to cause point discharge at numerous points on trees, buildings, etc. Even at the upwind mill, the basic V-pattern is much modified, probably by point-discharge effects from objects still farther to windward. There are also differences between the two mills which must have been caused by point space charges from the anemometer mast and from trees (the special point was down at the time). Under such conditions time intervals cannot usually be measured, as the two records do not correspond, and even when measurements are possible the time intervals have little meaning. Sometimes such time intervals give the impression of extremely high charge speeds and in Fig. 11 there are occasions when the downwind potential-gradient changes sign some minutes *before* that at the upwind mill. By the time the cloud had moved on another 900 m to the Laboratories, it appears that the positive space charge from the various points had been sufficient to make the potential gradient generally positive and therefore to produce negative space charge at points. This process further complicates the pattern of potential gradient.

Records such as Fig. 11 show something similar to Simpson's (1949) ' wave patterns ' but differ in that they are not symmetrical about the zero potential gradient. These patterns appear to indicate rhythmic changes in air motion, rather than in the basic electrical phenomena. It would be interesting to investigate in detail the relation between the potential gradient variations and the variations in wind at or near the discharging point.

The conclusion can be drawn that a potential-gradient pattern may be very considerably affected by point space charges. These patterns may give very little information about the electrical structure of the cloud and still less about the process bringing about charge separation within the cloud, since some of the cloud charges will be point space charges which have moved upwards.

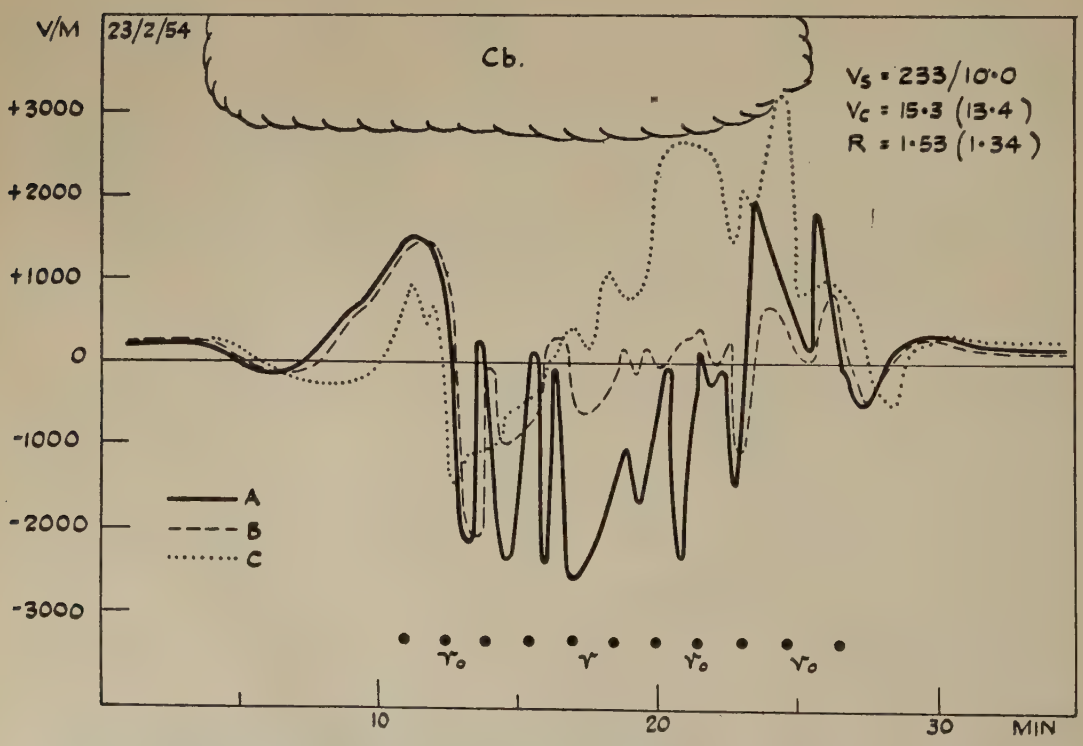

Figure 11. Simultaneous variations in potential gradient at three sites. Curve A: upwind potential gradient; Curve B: potential gradient 100 m downwind; Curve C: potential gradient 900 m downwind.

Davis and Standring (1947), in considering point discharge from a captive balloon, have attempted to calculate the potential-gradient change produced by point space charge by assuming that the liberated charge is carried downwind in the form of a line at constant height. Using this theory and taking the height of the point as 20 m and the length of the line charge as 100 m, the difference between the upwind and downwind potential gradients, each at 50 m from the point is

$$E_u - E_d = 3{\cdot}3 \times 10^3 \, i/V$$

where i is the point discharge current in μa, V the wind speed in m/sec, and $E_u - E_d$, the potential gradient difference, in v/m.

Some measurements of $E_u - E_d$ and i made in fairly steady conditions gave the following results. For low values of i/V, $E_u - E_d$ increased nearly linearly with i but its values were only about one third of those predicted by the simple theory. For higher currents the potential gradients were sufficient to cause discharge at other points besides the one through which i was measured. In this case $(E_u - E_d)/i$ would be expected to increase, as actually occurred. The failure of the theory to give an accurate prediction of the value of $E_u - E_d$ must be because of its over-simplification, for in reality the ions of the point space charge will spread out and not keep to a line, and some may be captured by falling precipitation.

11. Continuous snow and sleet

Measurements were made on six occasions during snow and sleet from layer clouds. One striking feature was the occurrence of moderately large positive potential gradients, possibly produced by the blowing of snow from the surface. On one occasion (Fig. 12) there were 18 reversals of sign during $2\frac{1}{4}$ hr of snowfall from nimbostratus cloud. It is this record which has been used for the analysis of time intervals in Fig. 2. The variations shown in Fig. 12 are comparable with Simpson's (1949) ' wave-patterns.'

12. Conclusions

The conclusions to be drawn are :

(a) In fine weather there are few and small variations in the potential gradient.

(b) In fair weather the variations are to be associated with the movement of wind-borne pockets of charge comparatively low in the atmosphere. Certain variations are believed due to charged locomotive steam whilst others appear to be related to convection patterns.

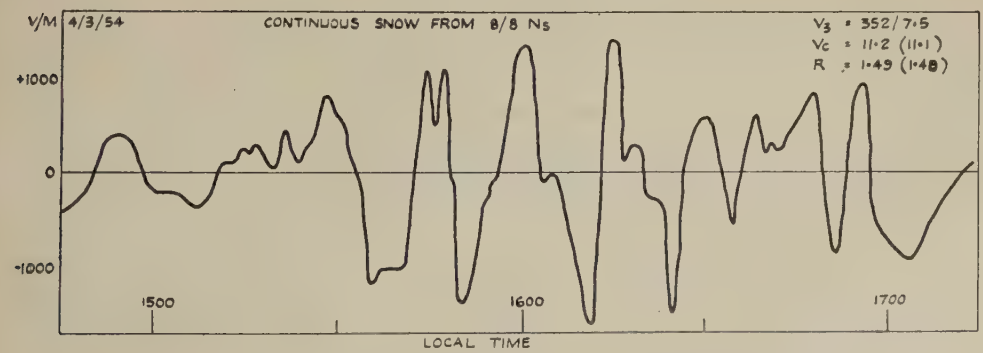

Figure 12. Potential-gradient variations during steady snowfall.

(c) With overcast sky the variations can be correlated with the cloud structure and may be interpreted as negative charges on the base of the cloud. These charges appear to vary in density according to the cloud structure, the more intense charges are found on the thickest portions of the cloud.

(d) With showers of not very great activity there is often a V-shaped pattern. This may be interpreted as a negative charge in the cloud base; often, however, the charges appear to lag behind the cloud. There is evidence for a positive charge in the heaviest rain, transforming the V-pattern into a W-pattern.

(e) With very active showers and storms the effects of space charge from point discharge destroy any pattern of potential gradient. The actual effect of the space charge is less than predicted by the simple theory.

13. Acknowledgments

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A regression method for estimating time changes of elliptically distributed winds

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Summary

The theory of a method, which is simple to apply, of finding regression equations for estimating time changes of wind is given. It is applied to two of the elliptical distributions found at Singapore, and its results are compared with those from two other methods. It is concluded that its use on such distributions is useful.

1. Introduction

Computation on the lines given in Scott (1956) on over five years' radar-wind observations at Singapore has shown that, up to 70,000 ft, practically all seasonal and a majority of monthly distributions are significantly non-circular. These distributions up to and including 60,000 ft were found to be nearly elliptical with the major axes of the representative ellipses generally not much inclined to the equator.

Durst (1954) and Ellison (1954) have described methods of obtaining regression equations for estimating time changes of wind; Durst's method requiring considerably less computation than the more general procedure of regressing one vector on another described by Ellison. Ellison and Walshaw (1955) found variance tensors which suggest that the winds they investigated are circularly distributed. They also found regression coefficients which suggest that their more elaborate computation does not produce results significantly superior to the method of Durst. The partial regression coefficients of each component on its successor are nearly equal while the other two are very small. Here then, circularity of wind distribution is associated with a tendency for the wind to change in such a manner that its vector ends on the line joining the vector ends of the mean and the previously observed wind (all vectors being drawn from a common origin). There appears no reason to expect a similar tendency in an elliptical distribution. However, it is natural to enquire whether some method of shortening computation and application similar to Durst's could be found to apply to the elliptically distributed winds over Singapore. This paper suggests and demonstrates such a method.

2. Theory

Ellison's method is based on the equations

$$E(u_1) = \alpha\psi_1 + \gamma\psi_2$$
$$E(u_2) = \epsilon\psi_1 + \beta\psi_2 \quad . \quad . \quad . \quad . \quad . \quad (1)$$

in which E denotes expected value of the quantity in brackets, u_1 and u_2 are the departures from the means of the W-E and S-N components of the wind to be estimated, ψ_1 and ψ_2 are those of the wind observed t hr before the estimate and α, β, γ and ϵ are regression coefficients. The equations upon which Durst's method is based are obtained by putting $\alpha = \beta$ and $\epsilon = \gamma = 0$. In considering the elliptical distributions of Singapore it is natural to enquire whether equations involving only the second of Durst's approximations might give better results than the use of both of them. Such equations would provide the desired reduction of computation and comparative ease of use. They are

$$E(u_1) = \alpha\psi_1, \qquad E(u_2) = \beta\psi_2 \; . \quad . \quad . \quad . \quad (2)$$

where α and β are to be selected so that

$$f^2 = \overline{(u_1 - \alpha\psi_1)^2 + (u_2 - \beta\psi_2)^2}$$

is a minimum. f is the root-mean-square error of estimate over the occasions used to determine α and β. Putting $\partial f^2/\partial\alpha = \partial f^2/\partial\beta = 0$ we obtain

$$\begin{aligned} \alpha &= \overline{u_1\psi_1}/\overline{\psi_1^2}, \quad \beta = \overline{u_2\psi_2}/\overline{\psi_2^2}, \\ f^2 &= \overline{u_1^2} + \overline{u_2^2} - \alpha^2\overline{\psi_1^2} - \beta^2\overline{\psi_2^2}, \\ R^2 &= (\alpha^2\overline{\psi_1^2} + \beta^2\overline{\psi_2^2})/(\overline{u_1^2} + \overline{u_2^2}) \quad . \quad . \quad . \quad (3) \end{aligned}$$

in which R is the correlation ratio of $\mathbf{u}$ on ψ.

To obtain the most accurate results from Eq. (3) it is necessary to evaluate all the quantities involved separately. However, if the observations available for computing $\overline{u_1\psi_1}$ and $\overline{u_2\psi_2}$ occur in comparatively long sequences we may develop as follows.

$$\sigma_{xt}^2 = \overline{(u_1 - \psi_1)^2}, \quad \sigma_{yt}^2 = \overline{(u_2 - \psi_2)^2}$$

in which σ_{xt} and σ_{yt} are the standard deviations, assuming zero means, of the changes in time t of the W-E and S-N components respectively. Whence

$$\begin{aligned} \alpha &= 1 - \tfrac{1}{2}\left(\frac{\sigma_{xt}}{\sigma_x}\right)^2, \quad \beta = 1 - \tfrac{1}{2}\left(\frac{\sigma_{yt}}{\sigma_y}\right)^2, \\ f^2 &= \sigma^2 - \alpha^2\sigma_x^2 - \beta^2\sigma_y^2, \\ R^2 &= (\alpha^2\sigma_x^2 + \beta^2\sigma_y^2)/\sigma^2 \quad . \quad . \quad . \quad . \quad (4) \end{aligned}$$

in which σ is the standard vector deviation and σ_x and σ_y the standard deviations of the W-E and S-N components, relative to the most acceptable mean, computed only from those observations which are used for computing σ_{xt} and σ_{yt}. When the mean has been found from a larger number of observations than that available for finding σ_{xt} and σ_{yt} the use of σ', σ_x' and σ_y' computed from all of them is likely to introduce errors in the computed values of α, β, f and R, though it will be seen that in the first example below they were not serious.

In what follows the equations analogous to Eqs. (3) and (4) developed from Durst's basic equations will be called (3D) and (4D) respectively while those analogous to Eq. (3) developed from Eq. (1) will be called (3E).

3. Application

Among 85 observations of wind at 03 GMT at 50,000 ft at Singapore made in March, 1951 to 1955, there were four sequences of 23, 7, 13 and 15 observations. The standard deviations of the components obtained from the 85 observations were $\sigma_x' = 24{\cdot}2$ kt and $\sigma_y' = 9{\cdot}9$ kt. The inclination to the equator of the major axis of the set of ellipses representing the distribution was $-4°$. The standard deviations of the components, relative to the mean of the 85 observations, computed from the 58 observations occurring in sequences were $\sigma_x = 24{\cdot}8$ kt and $\sigma_y = 10{\cdot}2$ kt.

TABLE 1. Regression coefficients, errors of estimate and correlation numbers from different regression methods applied to 50,000 ft wind distribution in March at Singapore

Equation applied	α	β	γ	ϵ	f (kt)	f_x	f_y	R
Eq. (3E)	+0·83	+0·24	−0·22	−0·10	17·4	14·4	9·7	0·77
Eq. (3)	+0·80	+0·30	0	0	18·0	15·0	10·0	0·75
Eq. (3D)	+0·73	+0·73	0	0	18·7	15·2	11·0	0·73
Eq. (4) } using σ_x and σ_y	+0·80	+0·26	0	0	17·9	14·9	9·9	0·74
Eq. (4D) } using σ_x and σ_y	+0·72	+0·72	0	0	18·6	15·1	10·9	0·72
Eq. (4) } using σ_x' and σ_y'	+0·79	+0·22	0	0	17·8	14·8	9·9	0·73
Eq. (4D) } using σ_x' and σ_y'	+0·71	+0·71	0	0	18·5	15·0	10·8	0·71

The full computation needed to apply Eqs. (3), (3E) and (3D) was made. Then the σ_x and σ_y above were used in Eqs. (4) and (4D). Finally σ_x' and σ_y' were similarly used. The results are given in Table 1, in which f_x and f_y are the errors of estimate (in knots) along the two component directions.

The contents of Table 1 confirm the accepted legitimacy of using equations such as (4D) instead of Eq. (3D). They also show that the use of variance parameters, computed from all observations used for finding the mean wind, produced results further from the best available values (the regression coefficients of the top three rows) than the use of those computed from only the observations used in computing wind change components.* Finally, they show that, operating on this elliptical distribution, Eq. (3) produced better results than Eq. (3D) though inferior to those from Eq. (3E).

That this latter inferiority is not universal is demonstrated by Table 2, where the results of application of Eqs. (3E), (3) and (3D) to another distribution are compared. This distribution was that of 91 observations at 03 GMT at 15,000 ft in October 1953, 1954 and 1955. Its parameters were $\sigma_x = 9{\cdot}4$ kt, $\sigma_y = 5{\cdot}4$ kt and the major axis of its representative ellipse was inclined at $-1°$ to the equator.

TABLE 2. REGRESSION COEFFICIENTS, ERRORS OF ESTIMATE AND CORRELATION NUMBERS OBTAINED FROM THREE REGRESSION METHODS APPLIED TO 15,000 FT WIND DISTRIBUTION IN OCTOBER AT SINGAPORE

Equation applied	α	β	γ	ϵ	f (kt)	f_x	f_y	R
Eq. (3E)	+0·79	+0·28	−0·13	−0·07	8·1	6·2	5·2	0·67
Eq. (3)	+0·79	+0·27	0	0	8·1	6·2	5·3	0·67
Eq. (3D)	+0·65	+0·65	0	0	8·5	6·3	5·7	0·63

A number of other elliptical distributions were examined by Eqs. (4) and (4D). In all cases the superiority of Eq. (4) over Eq. (4D) was demonstrated. It is of interest to record that equations analogous to Eq. (4) based on the assumption $E(u_1) = \alpha \psi_1$ and $E(u_2) = 0$ also produced results superior to those from Eq. (4D), though inferior to those from Eq. (4).

4. CONCLUSION

It is to be concluded that Eq. (4) might usefully be applied for estimating time changes of wind at Singapore. Regression equations are found and used with comparative ease by its application.

ACKNOWLEDGMENTS

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* In comparing (for example) the results of the second, fourth and sixth rows of Table 1 it is necessary to examine the values of R, not those of f. These latter values are much more sensitive to errors arising from the approximations involved in using Eq. (4) in place of Eq. (3). This is demonstrated by inserting the values of the regression coefficients in the expressions of f^2 and R^2 in Eq. (3), which operation produces the true errors and correlation ratios over the period investigated. The values of the error found by this method are 18·2 and 18·5 in place of 17·9 and 17·8 kt in the fourth and sixth rows of Table 1. The 18·6 and 18·5 of the fifth and seventh rows become respectively 18·9 and 19·2 kt. The associated values of R are unaltered.

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The zonal wind in the upper troposphere and weather at Singapore

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Summary

It is postulated that a relation may exist in certain seasons between the zonal component of the wind in the upper troposphere and weather at Singapore. On examination of observations it is concluded that such an association exists and is likely to be of use in short-range local forecasting.

1. Introduction

The northern summer and autumn (June to November) are the seasons of maximum low-level convergence and typhoon activity over the China Sea and the Philippines. Such activity is much less marked and less frequent, in all seasons, to the south-east of Singapore. Computation on over five years' radar-balloon ascents at Singapore has shown that these two seasons are also those of maximum mean east-to-west component of the 50,000-ft wind there (just below the mean tropopause); the means in both being about 40 kt, a value substantially higher than the means in the other two seasons.

Air moving towards the equator acquires a significant increase of velocity towards the west due to conservation of momentum. It seems not unlikely that, in the seasons concerned, when the zonal component just below the tropopause at Singapore is above the average, the air there may have recently diverged from the tops of bad-weather areas to the northeast. Air so diverging presumably has a strong negative rate of change of divergence as it moves from the immediate vicinity of the top of a bad-weather area, and, judging by the substantial fine areas usually adjacent to such areas, may attain a state of slight convergence. The possibility of this convergence persisting during subsequent motion suggests that above-average zonal components of wind just below the tropopause at Singapore may be associated with convergence there. Since such convergence would normally be associated with low-level divergence it appeared desirable to examine the time distribution of weather at Singapore during the seasons concerned, in conjunction with the daily values of the east-to-west component of the 50,000-ft wind. This paper describes some results of such an investigation.

2. Data used

The periods examined were August to November 1953 and June to November 1954 (daily radar ascents were commenced in August 1953). Daily 10-hr LMT (GMT plus 7 hr) values of the east-to-west component at 50,000 ft (u) were used. Estimates were made on days of no observation where the missing values could safely be assumed to be well above or well below average. There then remained six single values missing in the two periods. Each period of twelve hours, 00 to 12 hr LMT (a.m. period) and 12 to 24 hr LMT (p.m. period) was classified into one of three types, using the records of eleven autographic rain-gauges well scattered over Singapore Island. These periods were chosen because excessive rain at Singapore tends to occur in the middle of them during the seasons concerned; the ' sumatra ' squalls and thunderstorms around dawn and the convection rains in the late afternoon. The three types were :

D rain recorded at one gauge only or none at all,

N not a D or W period,

W 4 or 5 gauges reported rain, the sum of the falls being more than 3·00 in., or
6, 7 or 8 gauges reported rain, the sum of the falls being more than 1·50 in., or
9, 10 or 11 gauges reported rain the sum of the falls being more than 0·75 in.

The limits between N and W periods were chosen because they appeared likely to separate satisfactorily periods when excessive or widespread rain would have been obvious from those when falls were light or local. Rain at one station only was classified as a D period to take some account of the frequent occasions when isolated shower clouds of very small horizontal extent appear in atmospheres which could fairly be assumed to be generally divergent in low levels.

3. Results

Days upon which u was $\geqslant 50$ kt and had been so on the previous day, together with days on which u was $\geqslant 30$ kt and had been $\geqslant 50$ kt on the two days preceding, were found to have weather characteristics which differed significantly from those of the other days. Days in either of the above categories will be described as being α days. There were 88 α days in sequences of 2 (1), 3 (3), 4 (3), 5 (1), 6 (3), 7 (2), 9 (1) and 19 (1). Table 1 gives the frequencies of the various weather types occurring in the a.m. and p.m. periods of α days and their distribution over the whole period examined.

TABLE 1. Frequencies of weather types at Singapore, morning and afternoon, on α days and all days

Day Class	D	N	W	Total	
α days	43	25	20	88	a.m. periods
All days	157	75	73	305	a.m. periods
α days	71	11	6	88	p.m. periods
All days	156	82	67	305	p.m. periods

Besson's coefficient of persistence (Brooks and Carruthers 1953) of D p.m. periods over the whole period examined was 0·28, that over the α day sequences was 0·60 and that over non-α day sequences was 0·12. The sums of the α rows and 'all days' rows of Table 1 show a tendency for D periods to be more frequent during spells associated with above-average values of u. The breakdown given in Table 1 shows, however, that this tendency is entirely the result of a reduction of the relative frequencies of N and W p.m. periods; i.e., the effect of persistent above-average zonal component in the upper troposphere appears to be a significant damping out of afternoon convection with no apparent effect on the early morning 'sumatras.' An explanation of this unexpected result may perhaps lie in the fact that the latter storms almost invariably occur in predominantly westerly winds (up to at least 10,000 ft) of moderate speeds, and are likely therefore to be a delayed result of forced ascent over the mountain ranges of Sumatra : i.e., the impulse which starts the development of a 'sumatra' is probably a strong and systematic mechanical one and less likely to be damped-out by the existence of small divergence in the low-level flow than those which start the ordinary afternoon convection.

It is of interest to record that, over the period investigated, the parameter α would have had value as a short-range forecasting tool. The frequencies of D, N and W afternoons (12 to 18 hr LMT) were 169, 82 and 54 respectively, while those during sequences of α days were 78, 7 and 3.

4. Conclusion

It is concluded that above-average zonal components of the wind in the upper troposphere at Singapore are associated with fine afternoons during the seasons considered. It also seems likely that this association will be of use to short-range local forecasting.

Acknowledgments

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551.584.31 : 551.515.8

Temperature differences in the South Tyne Valley, with special reference to the effects of air mass

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(Manuscript received 25 October 1955, in revised form 15 March 1956)

Summary

Observations at three stations have been analysed to determine the temperature profile of a N-S valley. The relationships between them have been classified, their frequencies tabulated, and further analyses undertaken linking these relationships to air-mass types. An attempt has been made to correct the observed values for the period under review (October 1952 to May 1954) to the twelve years' basis used by Belasco for his work on air-mass frequency. It is concluded that air-mass properties have an effect upon the relationships other than those resulting from differences of wind speed and direction, radiation, and insolation.

1. Introduction

Throughout the period June 1952 to May 1954 observations were maintained at Alston in Cumberland to study the temperature profile of the lower 164 ft of the South Tyne River valley. Three stations provided the information; the Samuel King's School Climatological Station (*East*), and two thermograph records. The thermographs were housed in large Stevenson's screens and standardized thermometers were present to provide a regular check. These substations were at Nether Park (*West*) and at the riverside (*Valley*). To assist the comparison between the two sides of the valley, *West* was selected to be at exactly the same height as *East*

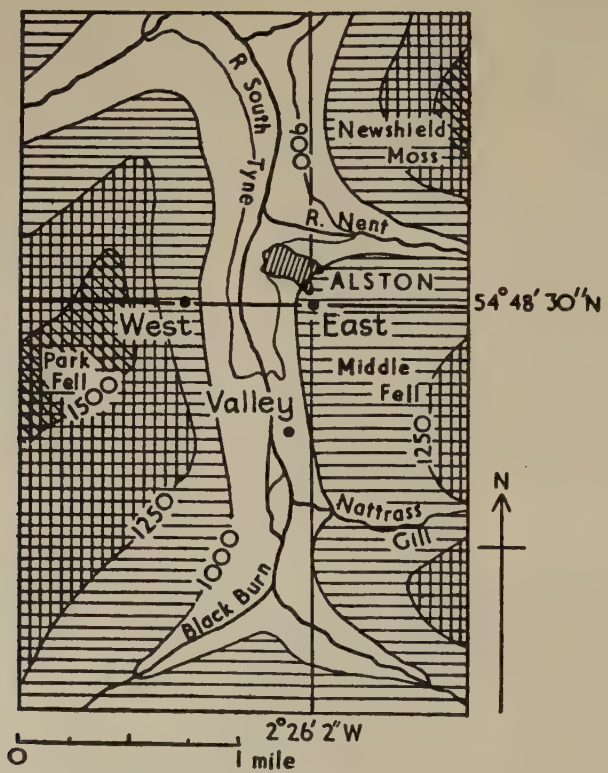

Figure 1. Map showing sites at which temperature observations were made near Alston.

(i.e., 1,074 ft a.s.l.). Three sites were tried for *Valley*; the one used throughout this work (i.e., after October 1952) was 70 yd from the river and 10 ft above it, at an altitude of 998 ft a.s.l. The earlier stations in the valley were considered unreliable, for the first was some 50 ft above the valley floor, and the second was so near the river that the temperature of the water might have affected results. The three stations are in a part of the valley which is of approximately north-south orientation, and consequently the durations of direct sunshine at each station are as nearly equal as it is possible to achieve in a valley. All three sites were over about 2 ft of loamy soil with a clay subsoil, and exposures were selected to conform with Meteorological Office requirements for climatological stations.

2. Temperature relationships

18 relationships between the values of the maximum and minimum temperatures at the three stations have been studied and the relevant data are shown in Table 1.

TABLE 1. Percentage frequencies of temperature relationships at Alston. Observed values refer to the period October 1952 to May 1954. The other series show values corrected by reference to Belasco's 12-yr means for air-mass frequency. Differences of temperature refer to variations of $\frac{1}{2}$°F or more.

Ref. No.	E = *East* W = *West* V = *Valley*	Per cent frequency of maximum temperatures		Per cent frequency of minimum temperatures	
		Observed	Corrected	Observed	Corrected
		of 468 obs.		of 459 obs.	
1	W = V = E	4	4	7	10
2	W > V > E	1	2	3	3
3	W < V < E	24	23	2	2
4	W < V = E	14	12	5	7
5	W = V > E	2	3	5	5
6	W < V > E	20	19	26	31
7	W > V < E	15	17	41	30
8	W > V = E	1	1	5	6
9	W = V < E	18	19	5	6
		of 471 obs.		of 465 obs.	
10	W > V	17	18	49	40
11	W = V	24	25	17	19
12	W < V	59	56	34	41
		of 512 obs.		of 515 obs.	
13	V > E	23	23	36	43
14	V = E	20	19	18	22
15	V < E	57	58	46	35
		of 490 obs.		of 481 obs.	
16	E > W	78	76	39	38
17	E < W	7	9	34	32
18	E = W	14	14	27	30
				of 188 obs.	
19	E > W	During occasions of full inversion (7) at night		34	—
20	E < W			39	—
21	E = W			27	—

The most frequent relationship is 16, which shows the maximum temperature at *East* to be higher than that at *West*; and the corresponding relationships 3, 12 and 15 also have a high frequency. This is to be expected in a north-south valley where the angle of insolation is more acute on the western than the eastern side during the afternoon when maximum temperatures are normally achieved. Of the frequencies for minimum temperatures the normal relationship based on lapse-rate, 6, has a high value, but the full inversion, 7, is much higher, though there is reason to suspect that 41 per cent is too high to be representative of the valley since the period of observation was more anticyclonic than is normal. This matter is pursued in greater detail in § 3 (*c*) and (*d*).

Significance might be attributed to the comparison of the minimum temperature relationships between 16 and 21, and especially to the change of high value from 17 to 20 and the reverse for 16 and 19. That *East* is cooler at night than *West* during the normal course of events is probably accounted for as a projection of the relationships of the daytime into the night. On the other hand the change to colder nights at *East* during nights of total inversions seems noteworthy, and may indicate a tilt of the inversion upper boundaries.

3. Effects of air masses

The basic variations due to effects of insolation and radiation having been considered, it seemed useful to re-analyse the relationships as consequences of different air-mass characteristics. Belasco's tabulations of air-mass frequency over the British Isles were accepted, as also were his definitions of air-mass characteristics (Belasco (1952). However, while the full classification of Belasco was used for the analysis, his 20 categories proved too numerous to be convenient and in consequence several were grouped together to give 7 for use in the final tabulations. These groups may be seen in the key to Fig. 2 with the references to Belasco's classification preserved.

To tabulate the relationships of *West*, *Valley* and *East*, the air mass at midnight was taken to be representative of the conditions producing the minimum temperatures, and noon was the time basis for the maxima. The air mass was determined by reference to the trajectories and characteristics displayed in the *Daily Weather Report* and a check was made by comparison with the 700-mb charts of the *Daily Aerological Record* for 0300 and 1500 GMT. By this means a series of four tables was constructed showing :

(*a*) *Actual frequencies* (*N*)

These tables are of little use in themselves as the number of observations varies from one group of relationship types to another.

TABLE 2. Percentage frequency of valley-temperature relationships with given air masses. Based on data of October 1952 to May 1954. Values are taken to the nearest whole figure so that additions do not necessarily give exactly 100 per cent.

				Minima					
Ref. No.	Am	Nm	Tm	Tc	Pc	Hg	Ho	F	UC
1	7	8	6	4	20	5	0	17	12
2	1	0	2	0	20	4	2	2	25
3	0	8	4	7	0	0	0	0	12
4	4	12	13	0	0	2	0	10	0
5	12	0	6	0	0	3	0	8	0
6	49	31	42	31	27	10	2	29	37
7	15	23	10	48	20	71	90	23	12
8	7	12	8	0	7	3	0	4	0
9	4	4	10	10	7	4	6	6	0
Actual frequency of air mass (midnight)	72	48	52	29	15	187	52	48	8

				Maxima					
Ref. No.	Am	Nm	Tm	Tc	Pc	Hg	Ho	F	UC
1	4	10	10	6	0	1	0	3	0
2	1	0	2	0	0	1	0	3	0
3	18	25	18	29	22	28	27	23	38
4	22	8	10	19	0	17	13	13	0
5	5	2	0	3	0	1	0	3	0
6	18	17	26	21	5	20	22	23	13
7	15	23	17	6	33	13	13	13	38
8	5	2	0	0	0	0	0	0	0
9	12	15	18	18	39	18	25	20	13
Actual frequency of air mass (noon)	77	53	67	37	18	138	37	70	8

(*b*) *Percentage values*

The percentage values $\frac{N}{A} \times 100$ where A is the number of occurrences of a given air mass, are a useful indication of the type of relationship to expect if an air mass can be reliably forecast. The calculated values for the group of relationships 1 to 9 are given in Table 2. Significant results for minima are the higher frequencies of type 6 with maritime air masses and fronts, and the even higher frequencies of 7 with the anticyclonic groups. For the maximum temperatures the clear skies of Hg lead to relationships 3 having the highest frequency, followed by 6 which reveals the normal lapse-rate tendency. The maritime groups (Am, Nm, and Tm) and fronts give similar results, but the relatively high value of 23 per cent of type 7 with Nm is curious, when contrasted with the figure under Am, as also is the difference of 22 and 8 per cent under these same air masses against 4.

(*c*) *Interpolated long-period values*

An attempt was made to eliminate characteristics which might not be typical had the period of observation been longer. For this purpose the long-period air-mass frequencies for Alston (*B*) were taken to be the mean of those given by Belasco for Kew, Scilly and Stornoway. The frequencies for Alston were then adjusted by dividing by *A* and multiplying by *B*. *A* varies slightly for each table analysed. In particular, differences occurred between the maximum and minimum tables because the one referred to noon and the other to midnight, whereas Belasco uses 1800 GMT. Table 3, produced by averaging all the 1952-4 values for Alston, gives an approximate idea of the values used. The most severe adjustments necessary were those for Nm, fronts and anticyclonic (Hg) groups.

TABLE 3. PERCENTAGE FREQUENCY OF AIR MASSES AT ALSTON.
A shows the average of observed frequencies between October 1952 and May 1954, and *B* the long-term frequencies (12 years) interpolated from Belasco's researches).

	Am	Nm	Tm	Tc	Pc	F	Hg	UC
≃ *A*	17	11	13	6	3	12	36	2
B	19	18	11	3	7	20	20	2

(*d*) *Correction to one year*

The tabulations described in sub-section (*c*) refer to sets of observations varying from 459 to 515 in number. They were reduced to a standard set of 365 to show the interpolated frequencies of relationship types with given air masses during one year. Fig. 2 shows the final results of the investigation.

As a final appraisal of the value of these long-term corrections the lists of the actual percentage frequencies of relationship types shown should be compared with those which result after the corrections described (see Table 1). No serious difference is to be found except between 6 and 7, 10 and 12, and 13 and 15 for the minima, where conditions are almost reversed. This, of course, is the result of the unusual frequency of anticylonic conditions during the period of observation.

4. RANGE OF VARIATIONS

The most interesting case of large temperature difference in the valley has already been described (Richardson 1954). During this night, 5-6 Mar. 1953, a temperature of 40·5°F was being recorded at *East* when at *Valley* it was 26°F. The difference between the minima for the night was 8°F. At 1100 GMT on 6 Mar. the relative humidity had the unusually low value of 16 per cent; on the same day Moor House produced its remarkable value of 7 per cent (Green 1953). On the night of 1-2 Feb. 1956 the temperature decreased from 10·5°F in the open air (the screen reading was 12°F) at *East* to 2·8°F when the instrument was carried by the observer to a point near *Valley* where there is an embankment for the Penrith road to cross the valley

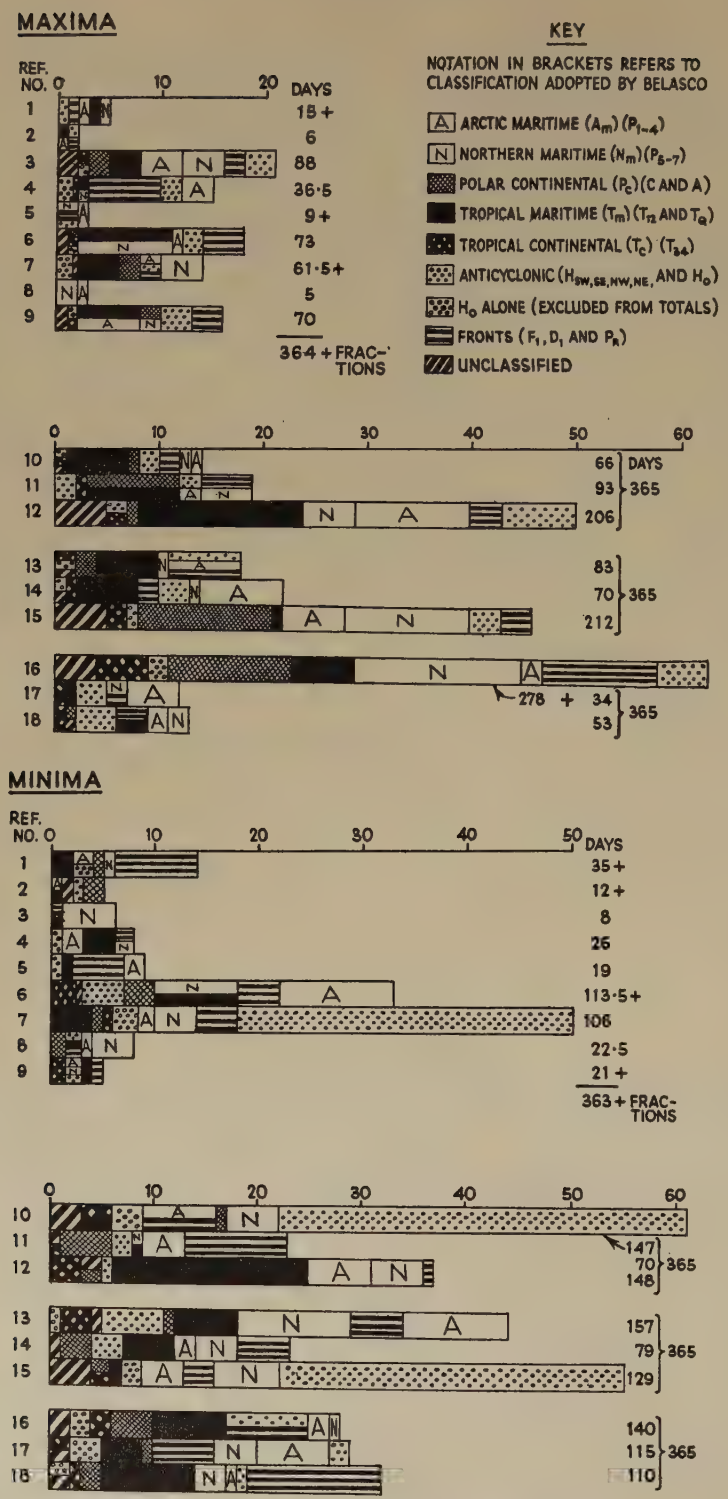

Figure 2. Probable frequency, in days during one year, of temperature relationships at Alston associated with different air masses. Air masses have been corrected for frequency by interpolation from Belasco's 12-yr values for Kew, Scilly and Stornoway. All values read from 0, and the total frequency of each relationship is given at the end of each line.

TABLE 4. FREQUENCIES OF GIVEN RANGES OF TEMPERATURE BETWEEN *West*, *Valley* AND *East*, ALSTON (OCTOBER 1952 TO MAY 1954) EXPRESSED AS PERCENTAGES OF THE TOTAL NUMBER OF OBSERVATIONS IN WHICH DIFFERENCES OCCURRED

MAXIMA (using 770 observations)

Range (°F) / Ref. No.	0·5	1·0	1·5 and 2	2·5 and 3	3·5 and 4	4·5 and 5	5·5 and 6
10	2·8	4·6	2·5	0·8	0·1	—	—
12	5·5	13·0	12·1	4·3	1·1	—	—
13	5·3	8·0	1·6	0·5	—	—	—
15	7·1	11·5	14·1	3·6	1·1	0·3	0·1
Totals	20·7	37·1	30·3	9·2	2·3	0·3	0·1

MINIMA (using 819 observations)

Range (°F) / Ref. No.	0·5	1·0	1·5 and 2	2·5 and 3	3·5 and 4	4·5 and 5	5·5 and 6	7	8	9	10
10	3·0	5·4	5·8	5·9	4·1	1·8	1·0	0·5	0·2	—	—
12	2·6	7·7	7·4	1·3	0·4	—	—	—	—	—	—
13	5·3	8·3	8·2	1·0	0·1	—	—	—	—	—	—
15	3·4	6·0	7·5	5·1	4·9	1·3	1·2	0·5	—	—	0·1
Totals	14·3	27·4	28·9	13·3	9·5	3·1	2·2	1·0	0·2	—	0·1

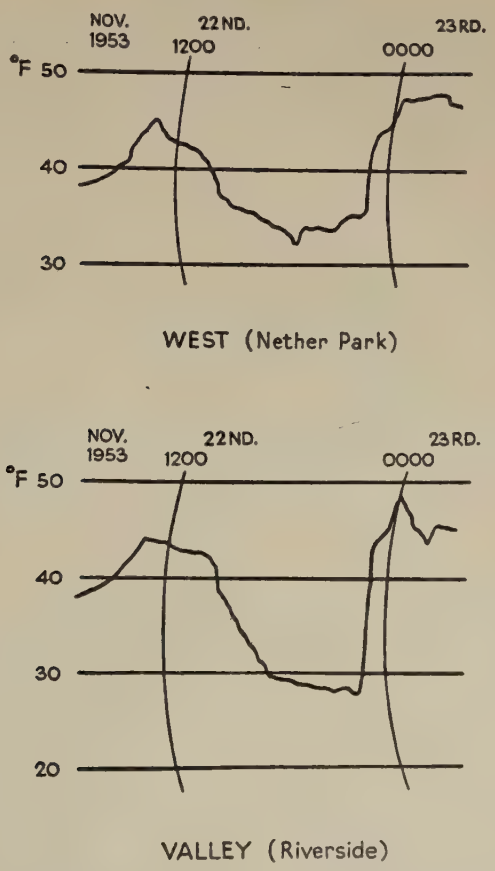

Figure 3. Reproductions of thermograph records for Alston on 22-23 Nov. 1953.

bottom. On 24-25 Feb. 1956 the screen minimum at *East* was 8·2°F but on an unscreened, although certified, thermometer at the same valley site — 4·7°F was registered (the 'grass' minimum at *East* was only — 2·8°F). Another interesting variation in the thermal properties of the two sites *Valley* and *West* occurred during the evening of 22 Nov. 1953. Radiative cooling had produced a well-developed inversion when a warm fast-moving air mass of tropical origin arrived at about 22 GMT and produced a rise in temperature of 17·5°F in one hour (and 21·5°F in two hours) at *Valley* compared with a rise of only 9°F in two hours at *West*. Adiabatic warming due to compression may well have contributed to this remarkable rise, in addition to the change of air mass (see Fig. 3).

For the more usual state of affairs Table 4 may be consulted. It is impossible to link the observations of *Valley* with the valley sides on any exact basis, but for clear nights the matter is treated theoretically elsewhere (Richardson 1955).

Conclusion

The present paper has attempted a quantitative analysis of some climatological phenomena of a valley, and it is considered practicable to forecast the temperature-relationship type provided there is a reliable forecast of the air-mass type. Although many of the frequencies shown in the tables and diagrams can be explained by reference to radiation, insolation, wind speed and direction, and cloud amount, these factors are in themselves aspects of air-mass study, and in addition there are several values which cannot be so easily explained and seem to be direct consequences of special air-mass properties.

Acknowledgments

The field work for this investigation was undertaken with a grant from the Scientific Activities Fund of the Royal Meteorological Society, and I wish to record my gratitude for this assistance, and also to the Department of Geography, Leeds University, and especially to Dr. S. R. Eyre, for kindness in providing advice and reference material.

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Correspondence

551.501.81 : 551.578.41

Factors influencing radar-echo intensities in the melting layer

By R. Wexler and D. Atlas

Mason (1955) has drawn some conclusions about the aggregation and orientation of melting snowflakes based on a simplified theory and measurements of radar-echo intensities in the melting layer. Although we agree that aggregation is an important factor in the melting layer, we believe that his analysis concerning the evidence for aggregation is incomplete. In addition, we believe that his conclusions concerning orientation are erroneous.

A factor that may frequently be important in the melting layer is the growth of the particles. Since Mason assumed an updraught, then growth must occur in the melting layer. Even without an updraught there will be growth in saturated air due to cooling of the air by melting snow (Wexler 1955), although this causes an increase in radar-echo intensity of only about 12 per cent.

Under steady-state conditions, the rain intensity R_w at the base of the melting layer is equal to the snow intensity R_i plus an increment ΔR due to growth, where

$$\Delta R = \rho u \, (\Delta q + \Delta L/\rho) \qquad . \quad . \quad . \quad . \quad (1)$$

where Δq and ΔL are the differences in specific humidity and liquid-water content respectively between the base and top of the melting layer. Assuming that the echo intensities are normalized with respect to distance and that fracture of the melted flakes is negligible, it may easily be shown that a corrected version of Mason's Eq. (1a), is*

$$\frac{I_i}{I_w} \simeq \frac{g_i}{5\nu} \frac{(v_w - u)}{(v_i - u)} \left(\frac{R_i}{R_w}\right)^2 \qquad . \quad . \quad . \quad . \quad (1a)$$

where g_i is a shape factor for dry snow, v_i and v_w are the terminal speeds of dry snow and raindrops respectively, u is the updraught velocity, and ν is the average number of dry flakes which comprise one raindrop.

To approximate to the factor $(R_i/R_w)^2$ assume that ρu is constant with height throughout the cloud. Then, neglecting vertical advection of liquid water,

$$\frac{R_i}{R_w} = \frac{q_{0°C} - q_t}{q_{3°C} - q_t} \qquad . \quad . \quad . \quad . \quad . \quad (2)$$

where q is the saturation specific humidity and the subscript t represents the cloud top and 3°C is assumed to be the bottom of the melting layer. Along the 10°C saturated adiabat we get $(R_i/R_w)^2 = 0{\cdot}66$ and $0{\cdot}61$ for a cloud top at $-$ 20°C and $-$ 15°C respectively. Cloud liquid water may, in addition, be carried up from below by the updraught and would tend to be depleted primarily by accretion in the melting layer (Atlas 1955a). A reasonable value of ΔL is $0{\cdot}2$ g m^{-3} depending largely on the distance between the cloud base and the melting layer. With this additional factor and the 12 per cent increase due to the cooling effect of melting snow, the ratios would be reduced to $0{\cdot}53$ and $0{\cdot}49$. Hence this factor may on occasion be quite appreciable.

Turning now to the question of orientation, Mason computes the shape factor g_m for the melting snowflakes from the simplified equation :

$$\frac{I_m}{I_w} \simeq g_m \frac{v_w}{v_m} \qquad . \quad . \quad . \quad . \quad . \quad . \quad (3)$$

He then computes $g_m = 5$ to account for $I_m/I_w = 7$, by using $v_w = 425$ cm sec^{-1} and the relatively high value of $v_m = 300$ cm sec^{-1}. It appears unlikely that such a large change in shape factor would be accompanied by such a small change in the fall velocity. The 5 to 1 reduction in g_m must occur upon the collapse of quite flat particles to raindrops, and such a collapse would undoubtedly be accompanied by an increase in fall velocity greater than the 1·4 used by Mason. Secondly, it presumes a relatively high change in fall velocity (by a factor of 4·3) from dry snow to the peak of the bright band, which would almost cancel out the change in the dielectric factor. The sharp increase in echo intensity at the upper portion of the melting layer would be ascribed by Mason to the shape factor, which is contrary to observation, as amplified below.

* Mason's Eq. (1a) is to be preferred to his Eq. (1b) since it is more accurate. From Eq. (1a) it is found that $I_i/I_w > 1$ for all $D > 1$ mm instead of 0·5 mm as he states.

Perhaps the most inconsistent aspect of Mason's inference that the melting particles scatter the radiation as horizontally-orientated oblate spheroids of diameter/axis ratio 4, is that such particles would scatter much less than equivolume spheres when viewed by a vertically polarized radar at low elevation angles. Below 10° beam elevation, for example, the particles would scatter less than 1/4 as well as equivolume spheres (Atlas, Kerker and Hitschfeld 1953) and this factor would wipe out any significant reflectivity difference between the melting band and the rain below. In other words, there would be no bright band at all at low elevations. Clearly this is at odds with the numerous RHI observations of the vertically polarized TPS-10 and CPS-9 radars, which generally show outstanding bright-band structures at low-elevation angles. Finally, it may be noted that observations by Newell (1956) indicate that in no case is the signal from the bright band greater by 1 db (a factor of 1·25) for horizontally polarized than for vertically polarized radiation, thus indicating a very slight orientation effect. However, an increase in reflectivity due to shape is possible when randomly orientated non-spherical particles become water coated (Labrum 1952, and Atlas et al. 1953). Deductions concerning the magnitude of this effect, show that they range from a factor of 1·2 to 3 based on measurements by Hunter (1954) and Newell (1956). Newell suggests an average factor of about 1·4.

In summary, we propose the following approximate equations for use in the interpretation of the melting layer,

$$\frac{I_m}{I_i} \simeq \frac{5g_m\,(v_i - u)\,\nu_1}{g_i\,(v_m - u)} \left(\frac{R_m}{R_i}\right)^2 \quad . \quad . \quad . \quad . \quad (4)$$

$$\frac{I_m}{I_w} \simeq \frac{g_m\,(v_w - u)\,b}{g_w\,(v_m - u)\,\nu_2} \left(\frac{R_m}{R_w}\right)^2 \quad . \quad . \quad . \quad . \quad (5)$$

where ν_1 and ν_2 denote the aggregation factors above and below the peak of the bright band respectively. Also b denotes particle break-up, an additional factor which Atlas (1955b) has indicated to be of importance in the lower portion of the melting layer where very large flakes may be present. In the light of the variable factors involved one can understand the wide range in the measured reflectivity ratios.

Measurements of the echo intensity have been made through the melting layer with a CPS-9 radar at Massachusetts Institute of Technology in widespread light rain during occasions when growth due to updraught in the melting layer could be neglected. Average values of I_m/I_i and I_m/I_w for 57 cases are given in Table 1, where I_i and I_w represent measurements at 1,500 ft above and below the peak of the bright band. Vertical resolution for these measurements was approximately 1,000 ft. In order to estimate the magnitude of the factors ν_1 and b/ν_2 we have made reasonable approximations for the mean values of the other factors in Eqs. (4) and (5). The average values of the shape factors are taken from Newell's results, referred to above. A value of v_i/v_m was chosen as 0·7 based on computations by Wexler (1955). The growth factor is assumed to be entirely due to the cooling effect of melting snow.

TABLE 1. EVALUATION OF THE PHYSICAL FACTORS IN THE MELTING LAYER IN WIDESPREAD RAIN

$\frac{I_m}{I_i}$	$\frac{g_m}{g_i}$	$\frac{v_i}{v_m}$	$\left(\frac{R_m}{R_i}\right)^2$	ν_1	$\frac{I_m}{I_w}$	$\frac{g_m}{g_w}$	$\frac{v_w}{v_m}$	$\left(\frac{R_m}{R_w}\right)^2$	$\frac{b}{\nu_2}$
10	1·4	0·7	1·05	1·95	5	1·4	3·3	0·9	1·2

In this evaluation aggregation of the dry flakes or crystals by a factor of about 2 is found for the upper portion of the melting layer. In the lower portion, break-up appears to exceed aggregation by a small fraction. It is a coincidence that the average aggregation factor ν_1 is about the same as that which Mason computed with greatly different ratios for the various terms. This is due largely to his over-estimate of the shape and velocity factors in the upper part of the bright band. It is to be noted also that the ratio I_m/I_i is very sensitive as to where measurements are made in the dry snow. Measurements made at 3,000 ft above the peak of the bright band give an average value of $I_m/I_i \simeq 25$, and another factor of 2 would then be ascribed to aggregation.

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11 April 1956.

Geophysics Research Directorate,
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Reply

By B. J. Mason

I agree with Wexler and Atlas that my Eq. (1a) could be multiplied by the term $(R_i/R_w)^2$ to take account of any growth of the snowflakes that might occur in the melting zone. However, I think this effect must generally be rather small (Wexler and Atlas's best estimate for this term, as given in their Table 1, being only 0·86), and in view of wide observed variations in the intensity distribution in the neighbourhood of the melting band, it is legitimate to neglect it.

Regarding my conclusions from the available radar observations on the orientation of melting snowflakes, I must admit that the data permit of other possible interpretations; the one given in my paper, when there were no reliable data on the polarization of radar echoes, seemed to me to be the most reasonable at the time.

If we take as a representative value of $I_m/I_w = 7$, where I_m, I_w are respectively the radar echo intensities from the melting zone and from the rain below, it follows that $g_m \simeq v_m/60$, where g_m is the shape factor of the melting flakes and v_m is their fall velocity. Wexler and Atlas question my choice of values for v_m and therefore g_m in this relation. My value of 300 cm/sec for the fall velocity of melting snowflakes was based on independent observations made both by Langleben (*Q.J.*, **80**, p. 174, 1954) and myself with ground temperatures of only about 35°F; I would therefore question Wexler's estimate (given in his Table 1) that the fall speed of a flake reaching the level corresponding to the peak of the melting band would be only 40 per cent greater than that for dry flakes. I would estimate the most probable value of g_m (corresponding to the peak of the melting band) to be about 3, although since the observed values of I_m/I_w lie between 2 and 14, values of g_m between 1·4 and 10 appear possible.

Our most important point of difference is upon whether or not melting snowflakes are horizontally orientated. At the time, it seemed to me intuitively reasonable that the disc-like melting flakes would, unless they were falling in a highly turbulent medium, present, on average, a maximum cross-sectional area to the vertically-pointing radar beam. It now appears that the polarization data suggest that there is only a rather small preference for such orientation; if this be the case, and I am not sure that the radar evidence at this stage is conclusive, then *randomly-orientated* water oblates of $g_m = 3$ would still require diameter/axis ratios of nearly 4.

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2 May 1956.

551.508.79 : 551.578.11

Stains produced by drops on filter papers

By R. T. Jarman

Fournier d'Albe and Hidayetulla (1955) stated that the relation between the drop and the corresponding stain diameter produced on an absorbent filter paper was given by the equation

$$D = a\,S^b \qquad . \quad . \quad . \quad . \quad . \quad (1)$$

where D and S are the drop and stain diameters, and a and b are constants. While a is said to vary with the paper used, b is said to be $\frac{3}{4}$ and this is stated to be expected from the geometry.

The relation between the drop and stain diameter produced on Whatman No. 1 filter paper has recently been measured for the following liquids: (a) tap water, (b) power kerosene, (c) a spindle oil, ASPA 21, (d) T.P. 5007, which is a 10 per cent solution of DDT in a mixture of 4 parts of power kerosene to 1 part of fuel oil, (e) a DNC solution consisting of 20 per cent Dinitro-ortho-cresol, 72 per cent Iranolin Kerosene Extract Bottoms and 8 per cent Iranolin 2241.

The water was dyed with Rhodamine G and the others with Waxoline Red. The drops were produced by a microburette (Lane 1947) and the diameter, found by measuring the volume and number emitted in unit time, ranged from 0·13 to 2·00 mm. The stains grew for 24 to 28 hr before the diameters were measured using a small magnifier. It was found that b was $0{\cdot}84 \pm 0{\cdot}02$.

Merrington and Richardson (1947), and more recently Furmidge (1954), have stated that the value of b is 1, if what they term the very small drops are excluded. From all these results it is clear that b does not take the value $\frac{3}{4}$ in all cases.

The geometry of the process is the flattening of a sphere. If we consider it is completely flattened we obtain (in effect) a uniform disc like a penny, and b takes the value $\frac{3}{4}$, while if on the other hand it takes up only a slightly oblate form the value of b is but little less than 1. The value taken probably depends on the period of spreading, the properties of the liquid and paper used, and the ratio of the drop diameter to the thickness of the absorbent papers, besides the impact velocity of the drops, etc. It is unlikely that a single universal value will be found for it.

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Discussion meeting on 'Measurement of water vapour in the stratosphere,' 15 February 1956

As already reported on p. 248 of this volume, a discussion meeting on 'Measurements of water vapour in the stratosphere' took place in the Society's rooms on 15 February 1956.

The following papers were read:

551.508.71

A method of increasing the range of the Dobson-Brewer frost-point hygrometer in jet aircraft

By P. GOLDSMITH

(See *Q.J.*, **81**, p. 607)

551.571.7

Some recent measurements of humidity from aircraft up to heights of about 50,000 ft over southern England

By R. J. MURGATROYD, P. GOLDSMITH and W. E. H. HOLLINGS

(See *Q.J.*, **81**, p. 533)

DR. G. D. ROBINSON: Could we be told something of the performance of the frost-point hygrometer in cirrus clouds? Does it in fact indicate saturation with respect to ice in these clouds? Is there a possibility that the visible deposit could lose molecules by surface migration?

DR. R. M. GOODY: Mr. Yarnell and I have obtained low-resolution solar spectra in the 6·3 μ region from a Mosquito aircraft flying at 35,000 ft. The band is very wide and we experienced great difficulty in fitting an accurate background. For this reason we doubt whether as little as one micron of water can be detected, as suggested by Drs. Houghton and Gates. This difficulty does not arise to the same degree with high-resolution grating spectra.

With regard to Professor Dobson's remarks about the drying of stratospheric air, I would like to draw attention to the fact that air will be rapidly dried somewhere above 70 km by photochemical decomposition. If the helium deficits measured below 20 km by Gluckauf and Paneth are real, then there must be some exchange of air between the ionosphere and lower stratosphere since the helium deficits could only develop in the ionosphere. A circulation which brings air from the tropical tropopause or the polar regions is therefore not the only possible explanation of the dry stratosphere in these latitudes.

I do not wish to urge this particular hypothesis but I would stress that the problem needs much more investigation, particularly by measuring stratospheric humidities at other latitudes.

MR. B. J. MASON: I should like to ask Dr. Brewer for a little more information on the appearance on the thimble of the frost-point hygrometer of the so-called 'glassy' ice at temperatures below about − 80°C. How is the glassy nature deduced – purely from its appearance and optical properties? Some experiments are now being carried out at Imperial College by Mr. Lisgarten in which ice is deposited from the vapour on to a cold metal (or formvar) surface inside an evacuated electron-diffraction camera. When the ice is deposited at temperatures above − 110°C it comes down in the normal hexagonal form; when deposited at temperatures between − 110°C and − 140°C its structure is definitely cubic; at lower temperatures still, it gives broad, diffuse diffraction rings indicative of an amorphous structure. These results are quite reproducible. Although the experimental conditions are, of course, rather different, the discrepancy between Dr. Brewer's − 80°C and our − 140°C is large.

MR. R. F. JONES: I wonder whether the American microwave refractometer (see, e.g., C. M. Crain, *Rev. Sci. Instr.*, **21**, No. 5, 1950) in conjunction with measurements of temperature would be sufficiently sensitive to enable measurements of vapour pressure at these high levels to be made?

(*Communicated*) : The instrument measures changes in radio refractive index of the ambient air. A change of frost point of 10°C from − 85°C to − 75°C with air temperature constant at − 73°C and at a constant pressure of 125 mb would lead to a change in refractive index of approximately 0·01 N units ($N = (n - 1)\,10^6$ when n is the refractive index). I understand, however, that at present the refractometer will measure changes of 0·1 to 0·2 N units so that a considerable improvement in sensitivity would be needed before it could be used at these heights. The potentialities of the instrument at lower levels should be borne in mind by meteorologists.

Dr. D. G. James : An analysis carried out by the Research Division at Dunstable of cirrus observations made from R.A.F. aircraft showed that of the 200 reports of cirrus cloud 18 were of extensive cirrus sheets in the stratosphere. On most of these occasions, the observing aircraft was itself above the tropopause, and reported a sheet of cirrus cloud still above the aircraft. Evaluation of the vorticity advection at 300 mb showed that 17 of these cases occurred when the sign of the advection was negative. Now, assuming little vertical motion at the tropopause a negative sign of vorticity advection at 300 mb implies a mean downward motion of air in the layer from 300 mb to the tropopause (and a mean upward motion in the lower stratosphere). I would be grateful if Mr. Murgatroyd or Dr. Brewer would suggest an explanation of the inconsistency of these cirrus reports with the frost points measured by the Meteorological Research Flight in the stratosphere.

Mr. R. J. Murgatroyd (*in reply*) : Replying to Dr. Robinson and Dr. James, Mr. Goldsmith and I (1953) have described the measurements made by the Meteorological Research Flight in and near cirrus clouds. It was found that the frost points were on the average usually about 3°F less than the temperature, suggesting that the air there was only about 80 per cent saturated. I agree that this seems improbable and there is a likelihood that at high levels the instrument may give readings which are a few degrees low. Several cases of cloud in the stratosphere were also noted in this report, some of which appeared to be haze. Up to the present the Meteorological Research Flight aircraft have never flown in cloud in the stratosphere which could be identified definitely as water or ice cloud. In fact there have only been two cases where the measured relative humidity exceeded 25 per cent. On the other hand it seems likely according to other observations (Barrett, Herndon and Carter 1949) that occasionally saturated layers may exist locally and of course mother-of-pearl clouds and perhaps noctilucent clouds are probably composed of water or ice.

With regard to Mr. Mason's remarks on the sublimation of ice on cooled surfaces the only evidence on this point from the Meteorological Research Flight's work is that it is relatively easy to observe frost points down to − 125°F or so, and on only one occasion up to date has an observer reported that he was unable to see the frost deposit on the thimble of the hygrometer at about − 130°F.

I am in full agreement with Mr. Jones's remarks on the microwave refractometer which is being used for flight investigations in the troposphere in America. At present the Meteorological Research Flight are able to measure temperature changes in the horizontal occurring in a few feet in the troposphere using a thermometer of very rapid response. When a refractometer becomes available, humidity changes on a similar or even smaller scale can be measured and more detailed investigations on the fine structure of the atmosphere will then be possible.

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Dr. A. W. Brewer (*in reply*) : With reference to Dr. Robinson's enquiry I would like to add to Mr. Murgatroyd's reply that I too found that the hygrometer seemed to indicate a small depression of the frost point in air which was obviously saturated. In my experience it was about 1°C (2°F) or slightly less, and occurred at almost all temperatures. All laboratory tests which we have made, however, confirmed that the fundamental basis of dew- or frost-point hygrometry is valid at all temperatures, and I regard this error as a peculiarity of the instrument, which does not invalidate the general picture we have seen today.

To Mr. Mason I would say that we only call an invisible deposit ' glassy ' for want of a better term (not that it is a very bad one). We have no evidence of its true state. It may have an ordered structure, rather than be of the nature of a liquid ' frozen in,' – but there is no doubt

from our evidence that above about − 90°C the equilibrium state of a deposit is of small crystals and that below − 65°C an increasing amount of time is required for the crystals to grow. These crystals have relatively large distances between them, and presumably an electron diffraction pattern would show the nature of the ' bare ' areas rather than of the small crystals. Below − 90°C the deposit is certainly invisible. We do not attribute this as necessarily due to a change in crystal pattern but to a decrease in mobility.

Cirrus clouds in the stratosphere are a problem to me as well as to Dr. James. The fact that the frequency of observation decreases with the skill of the observer is unfortunate, but some reports seem unquestionable. Except in very cold stratospheres very substantial upward movement would be required to produce cloud. There were several well-confirmed reports of cirrus clouds in the stratosphere in the summer of 1953 and at that time I thought that these clouds might be dust from some violent volcanic eruptions in Alaska.

Turning to the problem of the origin of the dry air of the stratosphere, I am sure that Dr. Goody is wise to raise the possibility of photo-chemical destruction. I considered this carefully some time ago, and on many grounds concluded that it was unlikely.

551.524.7 : 551.547.3 : 551.510.5 : 551.515

Variation of temperature in the troposphere and lower stratosphere

By J. K. BANNON and A. GILCHRIST

(Read 21 March 1956. See *Q.J.*, **82**, p. 58)

Dr. R. C. Sutcliffe (*President*) : A factual statistical paper of this nature may be valuable either because it is interesting and useful in itself or because it sets problems for explanation. Although the correlations are concerned with variations on the synoptic time-scale it must be realised that the explanation is only partly a synoptic-dynamical problem. To a large extent they derive from the latitudinal variations in the mean troposphere-stratosphere structure. The association of a warm troposphere with a high tropopause and cold stratosphere is a fact of world climate. This is a rather simple fact to state but so far as I know there is no simple explanation.

Dr. F. J. Scrase : I would like to show a slide illustrating some observations which confirm and extend the work of Mr. Bannon and Mr. Gilchrist. The slide shows temperature-height curves for some high-altitude soundings made at Downham Market, each curve being the mean of 3-6 soundings, all made at night. The soundings are classified according to season and tropopause height. The mean curves for the different tropopause heights in the left-hand part of the diagram show clearly the association between a low tropopause, cold troposphere and warm lower stratosphere, and vice versa, investigated by Mr. Bannon and his colleague. There is, however, little or no indication of any such association with conditions in the stratosphere above 20 km. From the right-hand part of the diagram it is very evident that above 20 km the seasonal variations are far more important than any which depend on the tropopause height. The diagram is included in a paper, now in press, in *Sci. Proc. Int. Ass. Met.*, Rome, 1954.

Dr. R. M. Goody : Mr. Bannon and Mr. Gilchrist have pointed out that at high levels the correlation with tropospheric conditions decreases, and the slides shown by Dr. Scrase indicate that it is very small at 30 km. This might very well be a radiative effect.

In a computation of radiative cooling it is possible to separate off a term caused by direct radiation to space from the terms involving other layers of the atmosphere. In the troposphere the term is small but it becomes more and more important as the height increases. This means that, the higher one goes, the less will the radiative equilibrium temperature be affected by conditions in other atmospheric layers and the more will it depend on the solar conditions.

Moreover, cooling to space is a very rapid way of disposing of heat and hence as we go higher we may expect the radiative equilibrium state to be approached with a shorter and shorter ' relaxation time.'

Thus in a general way we see why the higher stratospheric layers tend to become less dependent upon tropospheric conditions.

With regard to the changes in the region of the tropopause, I have no opinion to offer as to whether these are partly, wholly or not at all due to advection from other latitudes. However, if, as some speakers have suggested, advection is important, then a perfectly reasonable explanation is available. I was able to show in 1949 (*Proc. Roy. Soc.* A, **197**, p. 487) that a model atmosphere in which convection but not advection occurred would exhibit the decrease in stratospheric temperature with increasing ground temperature. This work needs revising in the light of better data, and cloud cover should be taken into account, but meanwhile it constitutes a reasonable explanation.

Mr. E. Gold : I suggest that the authors should examine the dependence of the correlation coefficients on the airstream. I found on rather inadequate data (*Geophys. Mem. Met. Off.*, No. 5) that the tropopause was at a lower level in a northerly airstream than in a southerly, for the same value of surface pressure. At 1,000 mb the difference in winter was about 2 km. Only when the variations with the airstreams or other fundamental meteorological conditions have been examined will it be practicable to answer the questions posed by the President.

(*Communicated*) : I have always felt critical of correlations between quantities connected by hydrostatic relations, as H and T of Table 1 are. For Larkhill for January the approximate hydrostatic relation between ΔH and ΔT is $\Delta H = 66\Delta T$. Using the standard deviations of Table 1 and an assumed unknown correlation r between H and T the regression equation is $\Delta H = 90\, r\, \Delta T$. Eliminating ΔH, ΔT gives $r = 0{\cdot}73$. For April, July and October the values of r are approximately 0·92, 0·89 and 0·92. Comparison of these values with the coefficients of Table 1 suggests that the latter are practically entirely due to the hydrostatic relation. I am inclined to believe that more of real meteorological significance would be learned from correlations with surface pressure of the height and temperature at the tropopause than from correlations between quantities related to a substantial degree by the hydrostatics. I found in 1912 (*Geophys. Mem. Met. Off.*, No. 5) approximate regression equations $\Delta H_c = 0{\cdot}07\ \Delta p$ and $\Delta T_c = -\ 0{\cdot}3\ \Delta p$ for mb, °C and km.

Mr. J. S. Sawyer : It would be interesting to know whether the correlations between stratospheric temperature and tropospheric conditions arise primarily from variations associated with individual depressions and ridges, or whether they are due to larger features of the upper flow, e.g., the long waves. This might be studied by examining the time scale of the disturbances upon which the correlations depend.

It is also rather surprising that the correlations involving T_{60} are highest in the summer when winds at the 60-mb level are light, so that advection and other dynamical effects must be small at this level.

Dr. R. S. Scorer : The correlation between cold troposphere and warm stratosphere suggests that the main dynamical influence at work is one which ensures that pressure gradients and winds are, on the whole, smallest at the ground; presumably this is some influence of the rigid boundary. Also at work are radiation influences which Dr. Goody mentions. With these two influences only would one expect a correlation less than is found ? If not there is no further explanation to seek. Could Mr. Bannon give us some idea how far we must go in seeking explanations for the correlations he finds ?

Mr. H. H. Lamb (*communicated*) : Mr. Bannon has pointed out how the correlations in this paper lend themselves to obvious geographical and seasonal sorting. This nice approach to the elimination of influences other than geographical and seasonal should greatly help all who read the paper to clarify their own ideas on the nature of the tropopause.

May we perhaps regard the tropopause in principle as the upper limit of mixing of the atmosphere by convection, vertical and horizontal and on all scales of magnitude, arising largely from the thermal behaviour of the earth's surface ? The greater horizontal air motions in the troposphere and the effects of cyclogenesis and anticyclogenesis complicate the issue by superimposing their own shifting pattern of disturbances of the tropopause level, through depression or uplift over limited regions, therewith involving the lower stratospheric air in the accompanying changes of level. Even the more straightforward vertical convection is not usually achieved in a single stage of mixing from ground to tropopause. Nevertheless, if one could eliminate the effects of cyclogenesis, anticyclogenesis and advection of régimes out of their own characteristic latitudes, it would presumably be found that the tropopause was carried highest in those regions and at those times of the year when vertical convection is strongest; the association between low tropopause and the polar regions of weak convection also has some general validity, but is interfered with by cyclogenesis, anticyclogenesis and radiation cooling of the ozone layer in the winter night. Can we then regard the coldness of the high troposphere and first part of the stratosphere over the regions of strongest convection as the regular result of convective

vertical currents overshooting their equilibrium level and forcing the development of a tropospheric adiabatic mixing layer relatively closer to the upper heat source in the ozone layer ? At some level higher up in the stratosphere, effects traceable to the heating pattern of the earth's surface would be expected to become secondary to those arising from the direct heating and cooling of the ozone layer.

An explanation along these lines, in spite of crude generalizations, might perhaps account for the stronger correlations of temperatures at 60 mb and 150 mb with events in the troposphere in the warmer regions and in the summer in higher latitudes.

In any case, I fancy I have put into words the sort of rather loose idea which many of us who are not experts on the subject have of the tropopause. How far are such ideas a permissible interpretation of the facts, or must we accept that at least in middle latitudes the complications due to the dynamics of pressure systems are ever present and really the main thing ?

Mr. J. K. Bannon (*replying on behalf of both authors, partly communicated*) : As the President has remarked, the association of warm troposphere with a high tropopause and cold stratosphere derives partly from the advection of these properties from lower latitudes. I regard the height of the tropopause as a function :

(a) of the degree of development (anticyclonic or cyclonic) taking place, and,
(b) of the air mass.

Mr. Sawyer (1954) estimates that (a) and (b) are roughly of equal importance. Mr. Gold suggests in effect that we should attempt to separate the effects of (a) and (b) and I agree this is desirable.

Mr. Gold points out that the correlations between H and T listed in Table 1 merely reflect the hydrostatic relation between temperature and height. These figures were not intended to do more than this and are not discussed in the text. Dines (1919) found a smaller correlation between surface temperature and height of the tropopause (0·68) than between pressure at 9 km and the height of the tropopause (0·84). For this reason we chose the height of the 300-mb pressure (roughly equivalent to the pressure at 9 km) rather than surface pressure as a parameter for analysis with respect to height (pressure) of the tropopause and temperature in the stratosphere. The contour chart of the tropopause surface resembles the 300-mb chart much more closely than the surface pressure map. Normand (1953) used the 300-mb chart for analysis of the distribution of ozone, and Bannon, Frith and Shellard (1952) found an association between features of the 300-mb chart and humidity in the upper troposphere and lower stratosphere. The 300-mb height is in my opinion a more useful parameter for this type of upper-air analysis than surface pressure.

Regarding Mr. Sawyer's question, I have the impression that the correlations between stratospheric temperature and tropospheric conditions arise primarily from variations associated with the usual succession of depressions and ridges and not because of variations on the scale of the long waves. I have not checked this numerically. I can suggest an explanation for the smaller T_{60} correlations in winter. Horizontal temperature gradients in the stratosphere are larger in winter than in summer; they also have different signs. In summer, other things being equal, advection from polar regions brings a warmer stratosphere and a lower tropopause, thus agreeing with the Dines model. In winter advection from polar regions brings lower stratospheric temperatures and slightly lower tropopauses, an effect contrary to the Dines model.

Dr. Scorer suggests that these correlations between the various upper-air parameters do not require further explanation. I agree with this to a certain extent, now that I know what the magnitudes of these correlations are; but I could not have forecast the coefficients before we made the calculations. Some refinement in the explanations is, however, still necessary. We need to know the separate contributions of advection and vertical motion to the relationships.

References

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551.577.34

Secular changes of tropical rainfall regimes

551.557.34 (94)

Secular variations of east-coast rainfall regimes

By E. B. KRAUS

(Read 21 March 1956. Presented by Mr. H. D. Hoyle. See *Q.J.*, **81**, p. 198 and p. 430)

Mr. D. J. Schove : The contrast between the climates of the 19th and 20th centuries shown by Dr. Kraus is very convincing. The method of cumulative residuals is dangerous unless homogeneity can be safely assumed. Nevertheless, my own results – part published in *Q.J.*, **76**, p. 147, 1950 – confirm the general tendencies shown by him. Indeed, the British Isles can for this purpose be regarded as a ' continent,' in as much as the contrast between our west-coast and east-coast trends (see Figs. 3 and 6 of my paper) is consistent with that found by Dr. Kraus for the continents. Thus, in parts of Lincolnshire, the standard period 1901-1930 is a rainfall minimum, although in the exposed west it is then much wetter than in the period 1861-1900. It would be most interesting if Dr. Kraus could prepare a world map showing the percentage change in long-period rainfall means between the nineteenth and twentieth centuries.

The change in rainfall-pattern is what we should expect from the known warming of the Arctic. We do not know whether temperature or pressure is the more fundamental parameter, but we would expect with the decreasing area of ice an associated fall of pressure in high latitudes, and this is confirmed by Arctic barometric data. The fall should be compensated by a rise of pressure in the Tropics and this again is confirmed. (Figs. 2 and 4 of my paper). The combined effects of pressure and pressure gradient lead in this way to drier conditions in low latitudes and on the eastern (leeward) sides of continents.

Why did the change come suddenly about 1895 ? My own results show a ' P-max ' about 1895 both in Europe and in S. Australia. In each region the P-max or belt of positive pressure anomaly progressed equatorwards so that westerlies suddenly became more dominant.

Pressure data available to me in 1950 in *World Weather Records* were not sufficiently homogeneous, but information since received from different parts of the World makes it possible to study the actual change in surface pressure and wind that took place about 1895. Thus, a map (unpublished) of the difference 1896-1905 – 1891-1900 shows a rise of pressure throughout Africa, the Indian Ocean and Southern Asia. The area of rise included Continental Europe (with the greatest rise off SW. Portugal) but not the extreme north. Near the Arctic Circle, and especially in SW. Greenland, there was a fall. There was also a fall in higher latitudes in the southern hemisphere (noted in Tasmania and southern S. America).

This pressure change explains the sudden rainfall change from wet to dry noted by Dr. Kraus for Cuba and the West Indies, Gambia, Bushire in S. Iran, Jaipur and Vizagpatam in India, the Madras summer rains, Queensland, SSE. Australia (Gabo Island), the semi-arid parts of western New South Wales, and Charleston and Hatteras in the Eastern United States.

The temporary reversal of the trend in the 1910's is best shown by charts of 5-yr, not 10-yr, moving means as it really consists of two spurts. Pressure maps show that there were two short reversals of trend about 1905 and 1915. The change about 1915 for instance is almost precisely the reverse of that described. Pressure fell throughout Australia (least in Tasmania), the Indian Ocean, S. Asia and Africa. A rise occurred in N. Asia and near the Arctic Circle, especially noted in SW. Greenland and Alaska. A corresponding rise could also be traced in the Southern Hemisphere in New Zealand and off SE. South America.

The general trend was resumed in the 1920's and slightly reversed in the 1930's. This general oscillation appears to be more universal and to have a longer period than the oscillations studied by Sir Gilbert Walker.

The sunspot cycle is still popular in interpreting tropical rainfall changes, but the data on sunspot and auroral activity (see, e.g., Fig. 1 in my paper in *J. Geophys. Res.*, **60**, p. 127, 1955), show that there is no change in magnitude between the two centuries. A contrast can of course be drawn between a period of declining activity from 1870-1900 and one of rising activity from 1900-1950, but this cannot seriously be offered as an explanation of the results of Dr. Kraus. Nevertheless, this evidence of historical aurorae and the numerous records of historical droughts and floods on the Eastern side of the Continents (see, Schove, *Met. Mag.*, **78**, p. 11, 1949) is now sufficient for all sunspot hypotheses to be put to a statistical test.

Mr. H. H. Lamb : The changes of rainfall in Argentina (as given by Schwerdtfeger and Vasino, *Meteoros*, IV, 1954) since 1900 differ from those in the regions studied by Dr. Kraus, but they are associated with changes in the general circulation which seem to fit in with some of his suggestions. Over the eastern half of the country rainfall has substantially increased, culminating near Buenos Aires with a 35 per cent rise between 1902-1911 and 1941-1950. Over a narrower strip close to the Andes the rainfall has decreased somewhat in the same period. The average pressure change from Buenos Aires to 53°S has decreased from about 12·4 mb in 1902-1911 to approximately 10·6 mb in 1941-1950 : this decrease must indicate a lessening in the extent of orographic shelter from the Andes with prevailing westerly winds and may mean either more frequent low index and meridional patterns, or less energy in the circulation.

There is great interest in Dr. Kraus's suggestion (which is in direct agreement with Riehl – *Weather*, **9**, p. 335, 1954) that a small change in the mean strength of the Trades, which is close to the critical value for smooth or rough sea surface, might lead to a great and discontinuous change in the evaporation and thereby in the energy available in the general circulation of the atmosphere. Since strong or weak Trades should, if this be so, mean respectively more or less evaporation and hence more or less energy in the atmosphere, both cases being apparently self-maintaining, one must consider what external factors might affect the evaporation independently enough to tip the balance in favour of the weak or vigorous cycles. The most obvious choice is variations in solar radiation. E. L. Deacon (*Austr. J. Phys.*, p. 209, 1953) recently examined the average daily maximum temperatures during the summer three months (DJF) at Alice Springs in the middle of the Australian desert and at three other stations in the interior, as the most direct indicator available of solar intensity, and found that in all cases the 10-yr running means had fallen rather steeply about the beginning of this century, latterly averaging 3° or 4°F (about 2°C) lower than in the 1880's and 1890's. Data available at Harrow suggest that there may have been parallel fall of 2° to 3°C in the average summer (JJA) maximum temperatures in the central Sahara (Timbuktu and El Goléa).

Prof. P. A. Sheppard : In the reading and discussion of this paper, the Society is, I believe, eavesdropping on the future – as well as on the past. For I expect meteorologists to become increasingly active in the field of climatic variation, not only in assembling and correlating the data on variation but in interpreting them in the context of the general circulation of the atmosphere or of its variations. Since we cannot yet be said to understand the general circulation itself, except perhaps in a very qualitative way, we are at least equally far from understanding its perturbations. But the study of the latter may well be highly profitable in the study of the former and both surely provide an exciting prospect in physical (including dynamical) climatology.

Whatever may be the certainty, or lack of it, in Dr. Kraus's findings – in a field beset by observational and statistical hazards – and in his hypothesis regarding the evaporation-precipitation cycle, we can expect the variable rainfall regimes he has examined to be a reflection of some features of the large-scale wind-field. (It would for example be very unlikely that a variation in rainfall arose solely from a variation in the content of water vapour, though the latter variation might repay investigation.) Since the Trades constitute a major feature of the general circulation, and are the medium of exchange for much of the energy which enters the atmosphere from below, it would be interesting to know whether there is any variation in their strength or direction to be matched against the variation of other climatic variables such as rainfall. The relations we seek are unlikely to be simple, perhaps not so simple as the relation suggested by Dr. Kraus, but he is likely to have considered the problem in some detail and we should be interested in his conclusions thus far, realising however that the problem is probably too big for one man to tackle. And if rainfall variation is related to wind we might anticipate some relation with meridional temperature gradient – Dr. Kraus's relation with the strength of upper westerlies perhaps ?

All in all, a large field to dig in the next decade or two. But what crops may we not reasonably hope to harvest ?

Mr. N. I. Cochrane : My work is largely concerned with the harnessing of large tropical rivers where changes of regime are of vital importance. I am not sure that my observations support Dr. Kraus. Suva in Fiji had an average rainfall of 102 in. per annum between 1884-1905. From 1905-1941 the average was 132 in., the reverse of Dr. Kraus's trend. In addition, I wonder if he has considered the implications of widespread changes of rainfall on the scale he suggests. The Lake Nyasa-River Shire complex, draining 49,000 mi^2 has a present mean rainfall of 48 in. per annum, only about 5 per cent of which emerges, on the average, as free water. If it had been the suggested 60 in. up to the end of the nineteenth century, we have data which suggest the mean free water in the complex would have been 4½ times its present value, at least

until some quite different vegetation was established. The Volta River in the Gold Coast drains 152,000 m^2. Its present rainfall is 43 in. and if the nineteenth century rainfall had been 54 in. the average river flow would have been about twice its present flow. The Nile is another low-yield river and even if we accept that the sudden change in its regime in 1899 was real and not instrumental it would only require a very small change in rainfall to produce a lowering of one-third in the river flow. Turning now to the effect of a generalized lowering of rainfall by 20 per cent on native population existing on rain crops, the effect would be catastrophic. In Africa the limit of intensive habitation is roughly the 40 in. mean annual rainfall line and Dr. Kraus's reduction would have meant the loss of one-third of the habitable area of the nineteenth century in Africa to natives existing on rain-crops. Such an adjustment could only take place by famine and death on a fantastic scale. Presumably, the effect in India and Asia would be even worse. I do not think that any of these major changes took place.

I take the opportunity to show two diagrams. The first shows the marked cyclic behaviour of Lakes Victoria and Nyasa with reference to the rate of change of sunspots, but not to the sunspots themselves. The second shows that the Smithsonian constant, the only measure of solar energy available to me, tends to be high when the rate of change of sunpots is high and vice versa. For Lake Nyasa, at any rate, the indication of this analysis is that there was less water in the complex at the end of the nineteenth century than there is now.

Mr. H. D. Hoyle (*in reply*) : Concerning Prof. Sheppard's remarks on the desirability of some measure of the strength of the Trades, there is some evidence in Lysgaard's ' Recent climatic fluctuations.' The pressure differences between Zi-ka-wei (E. coast of China) and Batavia (30-yr overlapping means are quoted), suggest that the winter N.E. Trades in this region decreased towards the end of the nineteenth century, and, after a period of minimum to about 1910, subsequently increased. In the summer, the pressure differences between these stations suggest that the SW. Monsoon in this region has decreased in intensity after the turn of the century. Whilst in some ways similar to the changes suggested by Dr. Kraus these pressure differences cannot be said to offer conclusive support.

In reply to Mr. Cochrane, a too world-wide interpretation of Dr. Kraus's results is not, I think, intended. For instance he emphasizes that in sub-tropical eastern regions of N. America the change is confined to the E. coast. In view of the large variations in rainfall which occur from year to year at almost any station, I find a percentage change of the order suggested by Dr. Kraus, in no way alarming and would suspect Mr. Cochrane's deductions more readily than Dr. Kraus's data. Many of the stations quoted by Dr. Kraus are in the ' arid fringe ' and he states that usually the greatest change is apparent in the dry season. A 25 per cent change in the dry-season rainfall of the arid fringe, whilst interesting meteorologically, does not represent a large amount of rain.

Dr. E. B. Kraus (*in reply, communicated*) : Similarities found by Mr. Schove between his results and mine are gratifying but may be difficult to interpret. The dynamic processes associated with rainfall at sub-tropical coasts and in the British Isles are often different. In particular, the east coasts will tend to be exposed to low-level on-shore winds in the sub-tropics. The opposite holds for the British East Coast. However, the upper winds in both cases would be mainly westerly and this might be a clue to the reported similarity. Variations in the strength of the 300 mb westerlies over parts of Australia have been shown (see Kraus, *Quart. J. R. Met. Soc.*, **80**, p. 59, 1954) to be probably associated with seasonal rainfall fluctuations. Mr. Schove's remarks about a sudden increase of the westerlies after 1895 are interesting in this connection.

Mr. Lamb mentions rainfall variations in the Argentine, which were different from those exhibited in Australian or North American East Coast records. I agree with his view that this is probably associated with the wind pattern in the lee of the Andes.

I am grateful to Mr. Lamb for pointing out the support which may be found for my hypothesis in Professor Riehl's article on the energy exchange between sea and air in the Trades. As regards the reported changes of maximum temperature in Central Australia and the Sahara, I would be at present reluctant to associate these with changes in direct solar radiation as received at the earth's surface. It is difficult to find any support for such an association in physical observations of solar radiation. Further studies are obviously desirable.

Mr. Cochrane states that in some tropical river catchments only about 5 per cent or less of the precipitation appears as stream flow on the average. This is, of course, true. However, I cannot follow his argument that a very small increase of rainfall must increase stream flow by an amount which should be more than twenty times as large. The argument seems to be based on the assumption that the contribution of additional rainfall to stream flow will be 100 per cent effective. This hardly can be the case. The more it rains and, in particular, the

longer the rainy seasons last (see ' Secular variations of tropical rainfall regimes ') the more will be lost by interception, evaporation during surface run off, and evapo-transpiration.

It is agreed that an increase of rainfall will cause a relatively larger increase in stream flow. This is a common fact of observations. If, for argument's sake, 5 per cent of precipitation appears as stream flow in a particular river catchment, a substantial increase of precipitation may cause perhaps six or seven per cent of the total to appear as stream flow. This is due to the relative decrease in evaporation associated with greater cloudiness and higher humidity. However, I am not aware of any evidence which would suggest the increase to be anywhere near as large as indicated by Mr. Cochrane.

In the paper on tropical rainfall it was estimated that the overall decrease was ' perhaps of the order of 20 per cent.' This figure finds some support in the published graphs and tables. Rainfall stations used for these were not particularly selected but represent a collection of all information, going back to at least 1880, published in *World Weather Records*. As most of the older stations had been established on islands or in coastal towns there may be some bias towards maritime conditions. However, the examples of Bogota, of inland Australian stations and of the Nile discharge suggest that the effect has not been smaller in at least some of the Continental regions though there were obviously large geographical differences. The exception constituted by the Monsoon region was specifically discussed. It is clearly brought out by Mr. Hoyle that the quoted figure of 20 per cent must not be taken to apply inflexibly to every region and station in the tropics as was assumed apparently by Mr. Cochrane.

Professor Sheppard's remarks allow us to see the whole subject in perspective. His comments on the study of the general circulation and its perturbations express my own view about the subject. Fluctuations in the general circulation will presumably cause the largest surface temperature variations in high latitudes where horizontal temperature contrasts are greatest. In low latitudes, temperature contrasts are comparatively small but rainfall varies greatly from place to place. Perturbation of the circulation would, therefore, be most apparent there as changes in the precipitation regime. That is the reason why fluctuations of the general circulation can probably be studied more easily from rainfall records than from temperature records in the tropics, and whilst the opposite holds in higher latitudes, this fact must limit the value and significance of a world map showing percentage changes of long period rainfall as suggested by Mr. Schove. That does not mean that such a project would not be of considerable interest.

Rainfall variation in the tropics and temperature variation in higher latitudes must affect the general circulation as they are affected in turn. The fluctuations of the evaporation-precipitation cycle and of the associated trade-wind circulation fall into this category. Whilst the general existence of such an effect is evident, its importance in the scheme of the general circulation and its fluctuations cannot be assessed without quantitative studies. The argument presented in the present paper is only qualitative. In a quantitative investigation changes in the meridional transport of heat and momentum, and changes of long-wave radiation, would have to be taken into consideration.

One very interesting suggestion in this connection was made by Mr. Woodrow C. Jacobs at Washington. He accepted the fact that a reduction of rainfall must have been associated with a reduction of evaporation. However, as all evaporation would tend to cool the ocean a reduction of the evaporation rate would necessarily change the heat balance of the ocean. The resultant increase in the sea temperature must prevent changes of the evaporation rate from becoming too large. The change in sea temperature is not necessarily observed most easily at places where the greatest variations in the heat balance occur. Transports in meridional currents may cause the observed temperatures change to be larger in high latitudes than in the tropics. Since Mr. Jacobs raised these points I have found in a 1956 University of Chicago publication by Riehl, ' Sea surface temperatures on the North Atlantic,' some possible confirmation of this argument.

Report of the Council

for the year 1955

(Submitted to the Annual General Meeting, 25 April 1956)

General

The Council are pleased to report a further year of progress. For the fifth year in succession there has been an increase over the previous year in the number of Fellows, Foreign Members and Associates of the Society; for the third consecutive year the total number of Members has exceeded 2,000.

A new local centre was formed at Winnipeg, Canada, during the year.

The second two-day Summer Meeting, consisting of scientific symposia, was held at Manchester University. Fuller details are given later in the report.

Two vacation courses were held in conjunction with the Council for the Promotion of Field Studies, one mainly for undergraduates at the Malham Tarn Field Centre and the other for sixth-form pupils at the Juniper Hall Field Centre. Both courses were very successful, and fifteen students were subsequently elected to membership of the Society.

Council

From 1 January to 30 September the constitution of the Council was as recorded in the 1954 Annual Report, except that Instr. Lt. T. R. Marshall resigned in June.

From 1 October 1955 the following Council held office :

President

Reginald Cockcroft Sutcliffe, O.B.E., Ph.D.

Vice-Presidents

Charles Kenneth Mackinnon Douglas, O.B.E., B.A.
Professor Gordon Manley, M.A., M.Sc.
Sir Oliver Graham Sutton, C.B.E., D.Sc., F.R.S.
Thomas Wilson Wormell, M.A., Ph.D.

Vice-President for Canada

Donald Patrick McIntyre, Ph.D.

Vice-President for Scotland

Sir Ernest Maclagan Wedderburn, O.B.E., D.L., M.A., D.Sc., LL.D.

Treasurer

James Strachan Farquharson, M.A., D.Sc.

Secretaries

Arthur George Forsdyke, Ph.D., D.I.C., A.R.C.S. (Librarian)
Howard Latimer Penman, M.Sc., Ph.D., F.Inst.P.
George David Robinson, Ph.D., F.Inst.P. (Editor)
Richard George Veryard, B.Sc.

Councillors

Charles James Boyden, B.A.
George Arthur Bull, B.Sc.
Henry Charnock, M.Sc., D.I.C., A.Inst.P.
James Marston Craddock, M.A.
Thomas Vivian Davies, M.Sc.
Ronald Frith, O.B.E., Ph.D.
Ronald Westlake Gloyne, B.Sc.
Frank Pasquill, D.Sc.
Professor Norman Pye, B.A.
Richard Segar Scorer, M.A., Ph.D.
Charles Edward Wallington, M.Sc.

In addition, Mr. Hubert Horace Lamb, M.A., was elected by Council in November to fill the vacancy caused by Inst. Lt. Marshall's resignation.

The Council met nine times during the year and there were thirty-seven Committee meetings.

Meetings

Eight Ordinary meetings were held in the Society's rooms during the year, including a joint discussion in June with the Challenger Society on 'Wind and waves.' There were also two additional meetings held in January and February: the first was a lecture on 'Chemical composition of precipitation as a function of atmospheric circulation patterns' by Professor C. G. Rossby, Honorary Member and Symons Medallist; the second was a joint discussion meeting with the Royal Astronomical Society, held in their rooms, on 'Hydrodynamical processes in meteorology and geophysics.'

In January, March and October, Mr. F. H. Ludlam gave a Popular Lecture on 'Clouds,' and in November and December, Mr. P. J. Meade gave a Popular Lecture on 'Fog and atmospheric pollution.'

The Annual General Meeting was held in the Society's rooms on 20 April when the Report of the Council and the Annual Accounts for 1954 were adopted; in place of the Presidential Address, a discussion meeting was held on 'The weather of 1954.'

The Symons Memorial Lecture on 27 April was given by Professor A. C. B. Lovell, Director, Jodrell Bank Experimental Station; his subject was 'Radio astronomy and the fringe of the atmosphere.'

On Empire Day, the Society and Messrs. Muirhead and Co. Ltd., were 'At Home' to visiting meteorologists from the Commonwealth.

The Summer Visit was to the College of Aeronautics, Cranfield, Beds., on 6 July by kind permission of the Warden, Mr. T. J. Palmer. Council wish to express their appreciation of a most enjoyable occasion.

Two-day Summer Meeting, Manchester, 21-23 July 1955

The second two-day Summer Meeting to be arranged by the Society was held at Ashburne Hall, Manchester University, by kind permission of the Warden.

The meeting included two scientific symposia, with an attendance at each of some seventy members and guests. The first symposium was on 'Some aspects of turbulence in the atmosphere' and the second on oceanographic topics. At the first symposium there were four opening speakers and at the second three.

A very enjoyable excursion was also arranged to Jodrell Bank Experimental Station, by kind invitation of Professor A. C. B. Lovell.

Council wish to record their sincere thanks to the Secretary of the Manchester Centre, Dr. Ruth Rogers, who was responsible for the local arrangements.

Publications

The inflow of scientific communications for the *Quarterly Journal* was slightly less than in the previous year. Sixty-nine papers were submitted for publication during 1955 as against eighty-two in 1954. Thirty-nine of the papers have already been accepted for publication. The average time between receipt of approved copy and publication in the Journal was about $3\frac{1}{2}$ months. The four numbers of the 1955 Journal together contained 662 pages, the same number as in 1954, this being the largest amount of material printed so far in any volume.

Council are greatly indebted to the successive Editors Dr. T. W. Wormell and Dr. G. D. Robinson, the Editing Committee and the proof readers Mr. E. L. Hawke and Cdr. C. R. Burgess, for their services in the production of the *Journal*.

The Society's monthly magazine *Weather* continued to maintain a high standard of production. Council are greatly indebted to the *Weather* Editing Committee for their valuable work.

Membership

Table A

	Fellows and Foreign Members			Associates	Students	Honorary Members	Total
	Annual	Life	Total				
Home Members							
31 Dec. 1954	1,369	174	1,543	29	102	18	1,692
Since elected	99	1	100	1	32	—	133
Re-admitted	1	—	1	—	—	—	1
Transferred	9	3	12	− 1	− 11	—	—
Deceased	− 9	− 4	− 13	—	—	− 1	− 14
Resigned, etc.	− 77	—	− 77	− 1	− 16	—	− 94
31 Dec. 1955	1,392	174	1,566	28	107	17	1,718
Canadian Branch							
31 Dec. 1954	48	6	54	275	—	1	330
31 Dec. 1955	46	6	52	298	—	1	351
Grand Total							
31 Dec. 1954	1,417	180	1,597	304	102	19	2,022
31 Dec. 1955	1,438	180	1,618	326	107	18	2,069

It will be noted that for the first time the Membership figures have been shown separately for Home and Canadian Members.

Deaths

The Council record with sorrow the death of Professor E. van Everdingen, Honorary Member, elected in 1925.

The following deaths are also recorded with regret, the dates of election being appended to the names : E. W. Bliss (Life Fellow, 1914), G. Bransby-Williams (1951), W. N. Cumming (1935), A. M. Gibb (1946), I. Jones (1935), B. W. Lee (1938), F. E. Lee (Life Fellow, 1920), Miss C. F. Marshall (Life Fellow, 1908), W. R. Pennell (1930), W. W. K. Pickard (Life Fellow, 1951), J. R. Pickett (1943), E. K. Watson (1949), and T. Williams (1951).

Library

In response to an appeal in *The Times* for old meteorological records, thirty replies were received. Of these, fourteen related to places where records were held, and to which access could be obtained on application; seven contained samples or extracts from diaries with the promise that further extracts would be supplied later. Three diaries received were of real value :

(i) A paper was prepared by Professor Manley and published in the *Q.J.*, October 1955 (p. 622) relating to the diary of Constantia Orlebar, Exton, 1786-1808.

(ii) Extracts are being made from a lengthy diary maintained by Samuel Winter, 1715-1786.

(iii) Extracts are to be made from a diary kept by David Wylie, Perth, 1823-1849.

In addition, several press enquiries resulted; articles were subsequently published in the *Manchester Guardian* and the *Birmingham Post.*

To help in the preservation of the contents of the library, the shelves in the Members' Room were glazed. It is hoped to extend the glazing to other rooms when funds permit. An additional cabinet for current journals was obtained for the Members' Room.

There has been a revised issue of the list of current publications received in the Library, and there is now ready a list of all the series of journals, memoirs, technical papers, etc., held in the Library, giving volume numbers and dates.

During the year loans included 322 books, 10 sets of lantern slides and 5 films. 79 enquiries were dealt with. Three lists of recent additions to the Library were issued.

Council wish to record their thanks to the successive Librarians, Dr. G. D. Robinson and Dr. A. G. Forsdyke, for their valuable work during the year.

Staff

At the close of the year the staff consisted of Mr. M. E. Shoenberg, B.A., Assistant Secretary; Mr. R. S. Read, I.S.O., M.A., B.Sc., A.R.C.S., F.Inst.P., Assistant Librarian; Mrs. A. Wallis, Accounting Assistant; Mrs. W. A. Steward, Publications Clerk and Editorial Assistant who, until 31 December 1954, carried out part-time duties as Publications Clerk, and Miss G. C. Jones, *Weather* Clerk. Miss A .F. Hayward resigned her position as Secretarial Assistant and was replaced by Mrs. V. Boulton, followed on 6 October by Mrs. J. G. Johnstone. Miss Y. M. Cullum, Junior Clerk, resigned and was replaced by Miss E. Crafts, as from 16 May.

Council desire to place on record their appreciation of the devotion to the Society's interests which has been shown by all members of the staff.

Branch and Centres

Canadian Branch

The Canadian Branch was expanded during the year by the establishment in December of a new Centre at Winnipeg.

Eight meetings were held in Toronto during the year, in addition to the Annual Business Meeting.

Membership in the Canadian Branch continued to increase and at the end of the year included 351 members.

The members of the committee were : President, R. C. Graham; Vice-President, W. L. Godson; Vice-President for Canada, F. W. Benum; Treasurer, J. G. Potter; Secretary, K. T. McLeod; Councillors-at-large, H. P. Wilson, Edmonton; M. Hardman, Winnipeg; W. F. Ganon, Ottawa; E. A. Barks, Moncton; G. H. Muttit, Goose.

Montreal Centre. Six meetings were held during the year, including a symposium. The meetings were well attended, the last meeting of the year having an attendance of about 100 persons.

The symposium, which took place in October during the visit of the delegates to the Fifth Conference of the Canadian Weather Service, consisted of two sessions. At the first seven papers were read; at the second session, two. The Officers were : Chairman, R. H. Douglas; Secretary, W. S. Harley; Treasurer, W. Hitschfeld.

Scottish Centre

Four meetings were held during the year at Edinburgh University and an excursion was made to a Weather Ship in Greenock Harbour in May.

On the occasion of the centenary of the foundation of the Scottish Meteorological Society, an exhibition entitled ' Scotland's weather 1855-1955 ' was organized by a Committee under Mr. T. Routledge and opened by Vice-Admiral W. A. G. Robson in the Royal Scottish Museum on June 6. During the period of the exhibition 6 June-2 July, there were over 14,000 visitors.

Sir Ernest Wedderburn, who was Secretary of the Scottish Meteorological Society at the time of its amalgamation with the Royal Meteorological Society in 1921, kindly consented to return as Scottish Chairman for the year of the centenary, and opened the 1955-56 session with

a lecture on the history and achievements of the Scottish Society, at which the President presided. After the lecture, Sir Ernest presided at a centenary dinner held in the rooms of the Royal Society of Edinburgh, and welcomed his guests, who included the President, some of the Officers of the Society and Professor C. T. R. Wilson

The Committee consisted of : Chairman and Vice-President for Scotland, Dr. A. E. M. Geddes from January to September, when Sir Ernest Wedderburn took office; Secretary and Treasurer, Mr. J. Paton, and nine other Fellows.

Manchester Centre

Six meetings were held in the Manchester Centre during 1955, including the first Annual General Meeting in October. The Centre acted as host to the Society at its two-day summer meeting held in Manchester in July, and a visit to Jodrell Bank Experimental Station was arranged. The Committee consisted of : Chairman, Professor P. R. Crowe; Secretary, Dr. Ruth H. Rogers; Assistant Secretary, Mr. N. K. Horrocks; Treasurer, Mr. G. Reynolds, and nine other Fellows, mainly from the Universities and the Meteorological Office in the area.

Representatives

The Society's representatives on other scientific bodies were :

Royal Society Committees
Gassiot Committee, President; *British National Committee for Geodesy and Geophysics*, Sir Graham Sutton (1952-57), T. W. Wormell (1955-60); *British National Committee for Physics*, F. Pasquill (1955-60).

Institute of Physics Committees
Journal of Scientific Instruments Advisory Committee, F. Pasquill (1955-56); *British Journal of Applied Physics Advisory Committee*, G. D. Robinson (1955-56).

Cambridge University Meteorological Committee
Sir Charles Normand.

Royal Astronomical Society
Geophysical Committee, G. D. Robinson.

Advisory Committee for the Meteorological Office, Edinburgh
Sir Charles Normand.

Awards

Symons Memorial Gold Medal for 1955

Dr. Reginald Cockcroft Sutcliffe, O.B.E., for his outstanding work in the field of synoptic and dynamical meteorology.

Darton Prizes for 1954

Home awards :

1st Prize : Dr. J. T. Houghton and Dr. A. W. Brewer for their paper ' A new radiometer ' (*J. Sci. Instr.*, **31**, p. 184, 1954).

2nd Prize : Dr. L. G. Smith, for his paper ' An electric field meter of extended frequency range ' (*Rev. Sci. Instr.*, **25**, p. 510, 1954).

Canadian awards :

1st Prize : Dr. D. P. McIntyre and Mr. R. Lee, for their paper ' Jet streams in middle and high latitudes ' (*Proc. Toronto Met. Conf. 1953*, p. 172).

2nd Prize : Dr. J. S. Marshall, for his contribution to his joint paper with Mr. T. W. R. East ' Turbulence in clouds as a factor in precipitation ' (*Q.J.*, **80**, p. 26, 1954).

The Awards were presented at the Annual General Meeting on 20 April.

Scientific Activities Fund

During the year the following grants were made :

Dr. H. U. Roll of Seewetteramt, Hamburg, in respect of attendance at the Joint Meeting with the Challenger Society in June.

The Gough Island Expedition, which took place from August 1955 to February 1956, towards the expenses of meteorological observations to be taken on Gough Island.

The Council for the Promotion of Field Studies towards the expenses of a newly-instituted vacation course in elementary meteorology for sixth-form students, organized jointly with the Council for the Promotion of Field Studies, and held in September 1955 at their Juniper Hall Field Centre, Surrey.

In addition, a grant was approved for payment in 1956 towards the travelling expenses of oversea delegates for attendance at a joint Conference with the Physical Society on ' Cloud Physics ' to be held in January 1956 in the Department of Meteorology, Imperial College.

Bequests

The Council gratefully acknowledge the bequests of a barometer (at one time used at 62 Camden Square by G. J. Symons) from Mr. T. C. Hawksley, and a microbarograph from Mrs. M. Morris in memory of her late husband.

Both instruments have been placed on loan at Field Centres of the Council for the Promotion of Field Studies.

Finance

The accounting system was changed last year from ' Receipts and Payments ' to ' Income and Expenditure ' and for the first time comparative figures under the new system are shown.

It will be seen that there is a small deficit this year as compared with a small surplus last year. This cannot, however, be regarded as altogether unsatisfactory in view of the continued trend of rising costs and the healthy position of the Society is reflected in its record membership and the flourishing condition of its publications as indicated elsewhere in the report. During the year approximately £700 was spent in redecorating the Society's offices in accordance with a scheme prepared by the Society's Consulting Architect.

£509 (as compared with £523 in 1954) has been received as a refund from the Inland Revenue in respect of Deeds of Covenant signed by Fellows in relation to their subscriptions. In view of the substantial benefits accruing to the Society under this scheme it is hoped that more Fellows will be encouraged to sign a seven-year deed for their subscriptions.

Council record with gratitude an allocation to the Society by the Royal Society of £500 from the Government Grant-in-aid of Publications Fund, with an addition of £600 for the Library.

GENERAL

INCOME AND EXPENDITURE ACCOUNT

EXPENDITURE

1954 £	£		£	£	£
		ESTABLISHMENT—			
	21	Architect's Fees		46	
	12	Water Rates		12	
	265	Cleaning Wages and Materials		275	
	22	Insurance		24	
	177	Lighting and Heating		206	
	190	Repairs and Maintenance		705	
	123	Depreciation of Furniture and Equipment		142	
810					1,410
		ADMINISTRATION—			
	2,275	Salaries		2,458	
	390	Contribution to Staff Pension Fund		349	
	148	Printing		165	
	109	Stationery		174	
	654	Postage, Carriage and Packing		501	
	79	Telephone		85	
		Audit and Accountancy Fees—			
	79	Royal Meteorological Society		105	
	21	*Weather* Magazine		31	
	26	Sundries		19	
3,781					3,887
		PUBLICATIONS, ETC.—			
	3,419	*Quarterly Journal*		3,228	
	272	Illustrations		328	
		Proceedings Toronto Meteorological Conference 1953—			
	1,425	*Printing*	—		
	146	*Reprints*	—		
	200	*Illustrations*	—		
	38	Postage and Packing	11		
	1,809		11		
	206	Reprints of Papers		377	
	7	Reviews	—		
		Monthly Weather Report	60		
		Less over-provided last year	38		
	100			22	
	670	Members' copies of *Weather*		934	
	102	Second-hand Journals		17	
6,585					4,917
		LIBRARY—			
	442	Salary		445	
	148	Books, Bindings, etc.		164	
590					609
		MEETINGS' EXPENSES—			
		London		157	
		Local Centres—			
		Scotland		40	
		Centenary Exhibition of Scottish Meteorological Socy.		106	
		Manchester		30	
137					333
100		PROVISION FOR BAD AND DOUBTFUL DEBTS			63
		GRANTS AND PRIZES—			
	16	*Buchan Prize*		—	
	20	*Grant to British Gliding Association*		—	
	25	Grant to Society for Promotion of Field Studies		25	
61					25
116		BALANCE BEING EXCESS OF INCOME OVER EXPENDITURE			—
£12,180					£11,244

FUND

FOR THE YEAR ENDING 31 DECEMBER 1955

INCOME

1954 £	£		£	£	£
		SUBSCRIPTIONS, ETC.—			
	5,086	Subscriptions		5,163	
	523	Income Tax recovered from Covenanted Subscriptions		509	
	105	Entrance Fees		113	
	13	Donations		16	
	5,727			5,801	
	—	*Deduct* – Adjustment for Lapsed Membership (By-law 40)		180	
5,727					5,621
		PUBLICATIONS—			
	2,211	Sale of Publications – current *Q.J.*		2,112	
	310	Sale of Publications – back numbers *Q.J.*		701	
	238	Reprints		457	
	262	Advertisements		317	
	1,284	*Proceedings Toronto Meteorological Conference 1953*— *Sales*, etc.		289	
	4,305			3,876	
	1,000	Grant from Royal Society (excluding Library)		500	
5,305					4,376
		INVESTMENT INCOME—			
	195	Interest on Government Stocks (gross)		260	
		236 Rent of Maisonette	236		
		104 *Less* Rates £88, Repairs *nil*, Legal Expenses £5	93		
	132			143	
327					403
		PHOTOGRAPHS—			
	18	Sale of Cloud Postcards		16	
		74 Clarke and Cave Copyright Fees	97		
		3 *Less* Reproduction Costs	6		
	71			91	
89					107
		LIBRARY—			
	600	Grant from The Royal Society		600	
	21	Sale of old Library Books		1	
621					601
		AMOUNT RECEIVED FROM OTHER FUNDS—			
	96	Life Composition Fund		92	
	15	British Rainfall Organization Fund (for administration)		15	
111					107
—		BALANCE BEING EXCESS OF EXPENDITURE OVER INCOME			29
£12,180					£11,244

GENERAL FUND
BALANCE SHEET AT 31 DECEMBER 1955

1954 £	£		£	£
		Accumulated Fund—		
	11,836	Excess of Assets over Liabilities at 1st Jan. 1955	12,516	
	564	Adjustment owing to change in basis of accounts	—	
	12,400		12,516	
	116	Excess of Expenditure over Income	29	
12,516				12,487
		Current Liabilities—		
		Creditors and Accrued Charges—		
	150	*Proceedings Toronto Meteorological Conference 1953*	—	
	1,428	Others	1,181	
1,578			1,181	
	412	Amounts received in advance— Publications	454	
	240	Subscriptions	330	
652			784	
243		Due to Bankers	220	
				2,185
£14,989			£14,672	

We have examined the above Balance Sheet and annexed Income and Expenditure Account which in our opinion respectively give a true and fair view of the state of the Society's affairs as at 31 December 1955, and of its income and expenditure for that year.

Spencer House,
South Place,
London, E.C.2. BALL, BAKER & Co.
29 March 1956. *Chartered Accountants.*

1954 £	£		£	£
		Fixed Assets—		
	5,375	Freehold Property – Estimated value (in 1925) of the Society's House, 49 Cromwell Road		5,375
		Furniture and Equipment—		

	At 1.1.55	Additions during year	Accumulated Depreciation	Net	
General	1,192	191	197	1,186	
Library	229	228	52	405	
	£1,421	£419	£249		1,591
1,314					

(No amount has been included for the value of the Library and the stock of publications)

		Investments—	
		Valued at market price 31.12.54—	
	924	2½% Consolidated Stock £1,400	924
	1,206	4% Consolidated Stock 1957 or after £1,250	1,206
	796	3½% Conversion Loan 1961 or after £900	796
	206	4% Funding Stock 1960-90 £200	206
	645	Government of Australia 4% Registered Stock 1955-70 £649.1.7	645
	2,030	Government of Australia 4% Registered Stock 1966-68 £2,000	2,030
5,807		(Market Value 31.12.55 £5,132)	5,807
	—	**Current Assets—** Stock of Stationery	50
		Debtors—	

£		£
650	*Proceedings Toronto Meteorological Conference 1953*	49
297	Subscriptions	345
490	Publications	390
1,407		784
100	*Less* Provision for bad and doubtful debts	150
1,307		634
23	Payments in advance	18
1,163	Deposit at Post Office and Cash in Hand	1,197
2,493		1,849
£14,989		£14,672

RECEIPTS AND PAYMENTS OF SPECIAL FUNDS FOR 1955

	Balance at 31.12.54	Income for 1955	Expenditure for 1955	Balance at 31.12.55
	£	£	£	£
Awards Fund	12	—	2	10
British Rainfall Organization Fund	146	157	62	241
Life Composition Fund	178	342	92	428
Napier Shaw Memorial Fund	106	74	—	180
Royal Society Reprints Scheme	22	2	18	6
Scientific Activities Fund	209	59	158	110
Staff Pension Fund	45	468 (a)	—	513
Symons Memorial Fund	54	31	56	29
	£772	£1,133	£388	£1,517

(a) Includes £350 received from General Fund, being 15% of salaries paid.

Examined and found correct.
29 March 1956.

BALL, BAKER & Co.,
Chartered Accountants

VALUATION OF INVESTMENTS AT 31 DECEMBER 1955

Market Value 31.12.54			*Market Value 31.12.55*	
£	£	*General Fund**	£	£
	924	£1,400 2½% Consols	791	
	1,206	£1,250 4% Consols	1,056	
	796	£900 3½% Conversion Loan 1961 or after	689	
	206	£200 4% Funding Stock 1960-90	185	
	645	£649.1.7 Government of Commonwealth of Australia 4% Registered Stock 1955-70	581	
	2,030	£2,000 4% Government of Commonwealth of Australia 4% Registered Stock 1966-88	1,830	
5,807				5,132
		Scientific Activities Fund		
	218	£250 3½% War Stock 1952 or after	191	
	88	£100 3½% Conversion Loan 1961 or after	77	
	1,030	£1,000 4% Trinidad Government Stock 1973-76	875	
1,336				1,143
		Staff Pensions Fund		
	708	£800 3½% War Stock 1952 or after	612	
	265	£300 3½% Conversion Loan 1961 or after	230	
	759	£1,150 2½% Consolidated Stock	650	
	868	£900 4% Consolidated Stock 1957 or after	761	
	515	£500 4% Trinidad Government Stock 1973-76	438	
3,115				2,691
		Life Composition Fund		
	528	£800 2½% Consolidated Stock	452	
	708	£800 3½% Conversion Loan 1961 or after	612	
	750	(a) £400 3% Defence Bonds	400	
	338	£350 4% Consolidated Stock 1957 or after	296	
	200	£550 3½% Defence Bonds Conversion issue	550	
2,524				2,310
		*British Rainfall Organization Fund***		
	654	£750 3½% War Stock 1952 or after	574	
	2,710	£3097.10.0 British Transport 3% Guaranteed Stock 1978-88	2,277	
	96	£100 4% Consolidated Stock 1957 or after	85	
3,460				2,936
		Napier Shaw Memorial Fund		
	310	£350 3½% Conversion Loan 1961 or after	268	
	386	£400 4% Consolidated Stock 1957 or after	338	
696				606
		Symons Memorial Fund		
	575	£650 3½% Conversion Loan 1961 or after	497	
	193	£200 4% Consolidated Stock 1957 or after	169	
768				666
£17,706				£15,484

(a) £350 3% Defence Bonds converted to 3½% Defence Bonds.

* General Fund also includes the Freehold of 49, Cromwell Road, S.W.7.

** British Rainfall Organization Fund also includes the Freehold of 62, Camden Square, N.W.1.

Examined and securities seen.
29 March 1956.

BALL, BAKER & Co.,
Chartered Accountants.

Proceedings at meetings of the Society

Symons Memorial Lecture, 11 April 1956

Mr. J. Paton, University of Edinburgh, delivered the Symons Memorial Lecture, his subject being ' The Polar aurora.'

Special General Meeting, 25 April 1956

Dr. R. C. Sutcliffe, O.B.E., President, in the Chair. A Special General Meeting of Fellows was held to approve an amendment proposed by Council, as given below, to the present wording of By-Law 86 :

> ' All the moneys of the Society not required to meet the current expenditure of the Society shall be invested in any of the Public Funds or Government Securities, or in any mode in which the Council are or shall be by law, in absence of special direction, authorized to invest trust moneys under their control.'

In the notice dated 12 March 1956 circulated to Fellows, to call the meeting, it was explained that owing to the continued depreciation of money, the interest derived from the Society's investments in Trustee Stocks has been considerably reduced in actual purchasing power.

It was therefore desired by Council to seek approval of an amendment to the By-Law which would enable part of the Society's investments to be made in any securities quoted in the official list of the Stock Exchange.

It was hoped by Council that this action would offset the depreciation by increasing the value of the Society's investments; action by the Society on these lines would be similar to that already taken by the Royal Society and the Royal Astronomical Society.

To give, therefore, effect to the desired change in the Society's investment policy the following amendment to the wording of By-Law 86 was submitted for approval :

> ' The moneys of the Society which are not subject to any specific trust purpose and are not required to meet current expenditure, may be invested in any securities in which Trustees are by law entitled to invest, or in the debenture or fully paid preferred, ordinary or deferred stocks or shares of any company quoted in the official list of any recognized Stock Exchange in the United Kingdom, provided always that no sum shall be invested during any year in non-Trustee stocks which shall exceed the amount approved by the Council and authorized by a Resolution passed at the preceding Annual General Meeting.'

The Resolution was proposed by the President, seconded by Dr. J. M. Stagg and, on being put to the vote, was passed by a large majority with only four dissentients.

Annual General Meeting, 25 April 1956

Dr. R. C. Sutcliffe, O.B.E., President, in the Chair. The minutes of the Annual General Meeting held on 20 April 1955 were confirmed and signed.

In presenting the Report of the Council for the year 1955 the President referred to the main items of interest. The programme of meetings held by the Society covered a wide field, including a lecture on ' Chemical composition of precipitation as a function of atmospheric circulation patterns,' by Professor C. G. Rossby and a joint discussion with the Challenger Society on ' Wind and waves.'

The Popular Lectures for sixth-form pupils, arranged in conjunction with the Local Education Authorities, had again been well attended.

The second two-day Summer Meeting, consisting of scientific symposia, held at Manchester, was very successful, and such meetings could now be regarded as a regular feature of the Society's activities. Special thanks were due to Dr. Ruth Rogers, Secretary of the Manchester Centre, for her help in making the local arrangements.

The Society's publications had maintained their high standard and thanks were due to the Editor, Editing Committee and the *Weather* Committee.

The membership of the Society continued to increase, although somewhat more slowly, and very good progress had been maintained by the Canadian Branch, with the formation of a new Local Centre at Winnipeg.

The President also referred with appreciation to the services of the Staff during the year.

After remarks by the Treasurer on the financial section in which he drew attention to the reasonably satisfactory financial position of the Society, the President moved that the Report be received and adopted and be printed in the *Quarterly Journal*. The motion was carried unanimously.

Consideration was then given to the Resolution submitted by Council and circulated to Fellows on 10 April arising from the new amended By-law 86 (approved at the Special General Meeting immediately preceding) as follows :

> ' that a sum up to £7,000 may be invested in non-Trustee stocks during the period of twelve months until the next Annual General Meeting.'

The Resolution was proposed by the Treasurer, who explained that the amount proposed for investment was equivalent to a little more than half the present invested capital of the Society, excluding the British Rainfall Organization Fund. Dr. J. Pepper seconded the motion and after discussion it was carried by a large majority.

The President then announced the following awards :

(a) *Buchan Prize*, in respect of the period 1951-1955 :

To Frank Henry Ludlam, D.I.C., for his outstanding contributions to the study of cloud physics as given in the following papers :

' The production of showers by the coalescence of cloud droplets ' (*Q.J.*, **77**, pp. 402-417, 1951), ' The heat economy of a rimed cylinder ' (*Q.J.*, **77**, pp. 663-666, 1951), ' The production of showers by the growth of ice particles ' (*Q.J.*, **78**, pp. 554-553, 1952), ' Orographic cirrus clouds ' (*Q.J.*, **78**, pp. 554-562, 1952); and, in collaboration, in the two papers :

' Bubble theory of penetrative convection,' by R. S. Scorer and F. H. Ludlam (*Q.J.*, **79**, pp. 94-103, 1953), ' Convection in the atmosphere,' by F. H. Ludlam and R. S. Scorer (Reviews of modern meteorology, No. 10, *Q.J.*, **79**, pp. 317-341, 1953).

Mr. Ludlam's researches have notably advanced knowledge on the incidence, form and behaviour of cumulus and cirrus clouds. He has elaborated theories for the generation of precipitation, rain, hail and snow, from shower clouds, so that the conditions under which any one form of precipitation is liable to occur are defined in some detail. His researches have been marked by excellent insight and reasoning and by a happy ability to turn visual observation of clouds to the clarification of the problems which they pose.

(b) *Hugh Robert Mill Medal and Prize* for 1956 :

To John Stanley Sawyer, M.A., for his outstanding contributions to the science of meteorology with particular reference to rainfall, as detailed below :

Mr. Sawyer has made important and original contributions to our knowledge of the distribution of rainfall associated with the passage of depressions and fronts across the British Isles, the results of his studies being given in two papers, *Quarterly Journal*, 1952 and *Meteorological Office Professional Note*, No. 118, 1956. Mr. Sawyer has also published two other valuable papers relating to rainfall, *Professional Note*, No. 96, 1949, in which it is shown that the cooling of air by rain may be an important factor in convection and *Professional Note*, No. 99, 1949, in which it is shown that the most important factor in producing downward air currents in rain is the cooling of the air by the evaporation of the rain.

Mr. Sawyer is the author of many other original papers in synoptic and dynamical meteorology and the high quality of the research conducted in the Napier Shaw Laboratory of the Meteorological Office owes much to his inspiring leadership.

The award of the Hugh Robert Mill medal is made in recognition of Mr. Sawyer's outstanding work on both the synoptic and physical aspects of rainfall.

(c) *Darton Prizes* for 1955 :

Home awards : 1st prize Mr. R. J. Murgatroyd, 2nd prize Mr. S. E. Ashmore.

Canadian awards : 1st prize Mr. Roy Lee, 2nd prize (jointly) Mr. R. Anderson, Mr. B. W. Boville and Mr. D. E. McClellan.

Dr. R. C. Sutcliffe, O.B.E., then delivered the Presidential Address, his subject being 'The moisture balance of the atmosphere.'

Dr. T. W. Wormell proposed a vote of thanks to the President for his Address and asked that he would allow it to be printed in the *Quarterly Journal*. Dr. E. T. Eady seconded the vote of thanks. The meeting was declared closed at 7.30 p.m.

Ordinary Meeting, 16 May 1956

The following candidates were declared to have been elected on 25 April by the Council :

Fellows

Bhalotra, Yash Pal Raj, M.A., Sudan Meteorological Service, Khartoum, Sudan.
Burns, Fred, 51 Dukes Avenue, Wealdstone, Middx.
Carlisle, Robert William, 165 Deerpark Road, Belfast, N. Ireland (*transfer from Student Membership*).
Chamberlain, Capt. John, 47 St. Nicholas Street, Ipswich, Suffolk.
Clayden, Thomas, B.Sc., 5, Tillmouth Gardens, Newcastle-on-Tyne, 4 (*reinstatement*, elected 1947, resigned 1954).
Cochrane, Noel James, B.Sc., A.M.I.C.E., Tregenna, 138 Tinsley Lane, Three Bridges, Sussex.
Coles, Ronald Lawrence, International Civil Aviation Organization, P.O. Box 978, Addis Ababa, Ethiopia.
Corbett, Walter Charles, 38 Wood Ride, Haywards Heath, Sussex (*transfer from Student Membership*).
Empire Cotton Growing Corporation, 12 Chantrey House, Eccleston Street, London, S.W.1 (*Nominee*, H. G. Farbrother, B.Sc., Cotton Research Station, Namulonge, P.O. Box 884, Kampala, Uganda).
Kennington, Christopher John, M.A., Yew Tree Cottage, 63 South Wexall, Bradford-on-Avon, Wilts.
Kestenberg, Morris, M.A., 30 Covington Road, Toronto, Canada (*transfer from Associate Membership*).
Scott, Peter Adamson, B.Sc., M.I.C.E., 49 The Ridgeway, Sutton, Surrey.
Stock, Jack, 46 Grove Street, Raunds, Northants.

Foreign Members

Burnett, Frank, B.S., 201 Belton Road, Silver Spring, Maryland, U.S.A.
Valladares, Alfredo Avila, Chief Meteorologist, Radio Aeronautical Office Mexicana, S.A. de C.V., Aeropuerto Central De Mexico, Mexico, D.F.

The following candidates were then declared to have been elected that day by the Council :

Fellows

Attwood, Sqn. Ldr. Arthur John, 5 Baldslow Road, Hastings, Sussex.
Lateef, Ahmed Mohamed Abdul, Meteorological Dept., Block No. 2, Frere Road, Karachi.
Morgan, Instr. Cdr. Maurice Richard, c/o Naval Weather Service, Admiralty, London, S.W.1.
Venugopal, Conjeevaram Govindaraja, 87 Warrender Park Road, Edinburgh 9, Scotland.
Woodburne, Alison Vera, Flat 2, 19 Holland Park, London, W.11 (*transfer from Student Membership*).

Foreign Member

Cadez, Dr. Marijan, Chief of the Aerological Observatory, Belgrade, Ljube Stojanovica 30, Yugoslavia.

The following papers were then read and discussed :

1. 'The temperature profile above bare soil on clear nights,' by J. V. Lake (*Q.J.*, **82**, p. 187).
2. 'Local temperature variations in the Reading area,' by M. Parry (*Q.J.*, **82**, p. 45).
3. 'On the space and time distribution of showers in a tropical region,' by C. M. Soane and V. G. Miles (*Q.J.*, **81**, p. 440), presented by Dr. A. G. Forsdyke.

THE NAPIER SHAW MEMORIAL PRIZE

Award of the first Napier Shaw Memorial Prize

March 1954 was the centenary of the birth of Sir Napier Shaw, first Director of the Meteorological Office, first Professor of Meteorology at Imperial College, President of the International Meteorological Organization, and President of this Society.

A fund was established by the Society to commemorate his unique position in meteorology and it was decided to award a prize of £100 at intervals of not more than three years for an original essay on a set subject.

The first competition was announced in April 1954 (*Q.J.*, **80**, p. 288), the subject being ' The energetics of the general circulation', and was closed on 1 March 1956.

In June 1956 the adjudicators recommended the award of this prize to Dr. Norman A. Phillips of the Institute of Advanced Study, Princeton, U.S.A. for his essay ' The general circulation of the atmosphere : a numerical experiment,' which had been published in the *Quarterly Journal* (**82**, p. 123) and this was announced by the President at the Ordinary Meeting of the Society on 20 June 1956.

The adjudicators also commended the excellence of the entry ' On the dynamics of the general circulation ' by Professor Robert G. Fleagle, of the University of Washington, Seattle, U.S.A.

Reviews

Introduction to dynamic meteorology. By H. A. Panofsky. The Pennsylvania State University, 1956. Pp. viii, 243; 60 Figs. Np.

Rapid progress has been made in dynamical meteorology since the Second World War, and the application of electronic computers to the dynamical aspects of forecasting has revolutionised methods of thinking regarding the dynamical problem. As a result much of the theory and experience which is now the accepted basis of methods of numerical forecasting and research is not to be found in the standard meteorological text-books.

The present book is therefore very welcome. It is perhaps not so complete as might be expected from its title, for no attempt is made to give full derivations of the basic dynamical equations of meteorology, nor of the special forms used in numerical forecasting. However, all the important equations and results are given and their significance fully explained. This treatment has the advantage that the book can be read and understood by anyone with a mathematical background of first-year university level. Very recent work is described, even some which has not yet been published.

The book has been written with the requirements of numerical forecasting in mind throughout, and will form an excellent introduction to meteorologist or non-meteorologist coming fresh to this field. The specialist will find little which is novel and it is doubtful whether he will find the book complete enough for reference purposes.

A chapter is included on perturbation methods, a useful review of the results achieved by linearising the equations of motion. Chapters are also devoted to turbulence and to theories of the general circulation, but these are little more than summaries of some aspects likely to be of interest to the would-be numerical forecaster.

The book is excellently produced with very few minor blemishes, although, perhaps, the confusion over coordinate systems on p. 34 may be mentioned. Here the x-axis is taken as directed everywhere towards the east, whereas elsewhere in the book the more usual convention of a straight x-axis on a suitable conformal map is used.

J.S.S.

Meteorologicheskiy Slovar. By S. P. Khromov and L. I. Mamontova. Leningrad (Gidrometeorologicheskoe Izdatel'stvo), 1955. Pp. 455; Figs., Tables. 18·9 roubles.

This glossary is stated in the preface to contain the terms used in contemporary Russian meteorological and climatological literature and in the conversation of meteorologists. The number of words defined probably exceeds 3,000; many are defined merely by reference to synonyms.

It has short descriptive articles for many words. The articles are briefer than those in the *Meteorological Glossary* of the Meteorological Office. It is well illustrated with charts and figures. A number of terms from physics, hydrodynamics, hydrology, astronomy, etc. of frequent occurrence in meteorological literature are included.

The book is on the whole up to date. For example, the jet stream, the sodium content of the upper atmosphere, rocket methods, Richardson's number, the formation of raindrops by coagulation in ' warm ' clouds, and artificial precipitation appear in their due places.

The authors admit in the preface that omissions may be found. A subject which the reviewer could not find in the book was the terminal speed of drops. Though Rossby's long-wave formula and the ' Method of trajectories ' (Rossby) – the method of constructing the future map of high-level isobars or isohypses of isobaric surfaces by theoretical calculation of the speed and direction of the wind – are given, ' constant vorticity trajectory ' is not defined as such. Such terms as ' barotropic model ' used in the western literature of numerical forecasting do not appear, presumably because they are not used in the U.S.S.R.

The terms used in Multanovsky's long-range forecasting method and in Taborovsky's method of upper air analysis are naturally described. The term ' metel,' blizzard, appears but not the word ' purga ' for the same phenomenon; possibly it has gone out of use.

It was noticed that the world maps of prevailing winds for January and July are identical, except for omission of the intertropical front, with those in the third edition (1938) of the *Meteorological Glossary* although only the second edition is included in the bibliography and though B. P. Alisov, the authors' colleague at the University of Moscow, very recently in his *Course of climatology* published more detailed maps including an indication of degree of persistence of the winds.

No factual errors were noted. The book is indispensable for all users of Russian meteorological literature and study of it would be an excellent way of acquiring a technical vocabulary by any meteorologist who had learnt some Russian.

G.A.B.

Physik der Sternatmosphären. By A. Unsöld. Berlin (Springer), 2nd Edition, 1955. Pp. x, 866; 287 Figs., 106 Tables. 168 DM.

Students of the first edition of Unsöld's *Physik der Sternatmosphären* will know what a significant book it was. Because of rapid advances in the 18 years since it appeared, the second edition has had to be an essentially new and far longer book. There is a new section on the emission of radio waves and cosmic

rays by the sun and stars; also the discussion of subjects treated earlier is altogether transformed. The author's reputation guarantees the excellence of the book; my only criticism is that on some topics it is a little too encyclopaedic. But the excitement which the first edition aroused is still to be felt, particularly in regard to the chapters on the sun.

Meteorologists will want to know what common ground the book reveals between their own subject and that of stellar atmospheres. Two-thirds of the book is concerned with radiation. General radiative processes are probably better treated in astrophysics books than anywhere else, and Unsöld's account of radiative transfer and the broadening of spectral lines is useful in understanding any problem of radiation. But his applications of general theory are mainly to isolated absorption lines, whereas in the earth's atmosphere band absorption is more important.

Unsöld devotes a number of sections to turbulence, mainly that due to thermal instability. He uses mixing-length concepts, and briefly discusses the turbulence spectrum. Stellar turbulence differs from terrestrial in that turbulent velocities in stars are often not small compared with sound velocities, so that shock waves have to be considered; also the turbulence is frequently affected by magnetic fields, the closest analogy in the earth's atmosphere being in ionospheric motions. Thus a meteorologist's main interest will be in the contrasts between the astronomical discussion of turbulence and his own.

All geophysicists are especially interested in the sun because of what it sends us, and Unsöld's account of solar phenomena is peculiarly good. Solar flares, leading to aurorae, magnetic storms and radio fade-outs; the hot corona, with its effects on the ionosphere; bursts of cosmic rays and radio noise from the sun; all these, and many others, are fully discussed. And even though the cyclonic theory of sunspots is no longer seriously regarded, the study of solar ' weather ' still has a real interest.

T.G.C.

A Mariner's Meteorology. By C. G. Halpine and H. H. Taylor. New York (D. Van Nostrand Company) 1956. Pp. xi, 371; Figs., Tables; Maps. $8.00 (42s.).

This book is written by two American Naval Officers who claim to have compiled it primarily as a standard work of reference for the teaching of meteorlogy to future officers at the U.S. Naval Academy. Its appearance is an event of significance for all who are concerned with practical applications of meteorology in maritime affairs, or who are responsible for making theoretical advances in marine aspects of the subject.

The authors have succeeded in producing a volume which is entitled to take its place among many similar text-books prepared in the United States as regards high quality of letter-press, illustrations and layout. A synoptic meteorologist is sure to feel at home in nearly every portion of the text. While the book will doubtless be of value to the Naval Officer, as a source of information for merchant seamen and yachtsmen it has a rather limited value. The readers to whom the authors address the book in their Preface should surely gain a fair grasp of the basic principles and methods of present-day forecasting, as practised at any main weather centre, after working through this volume. But it is more doubtful whether they can here gain much insight into the problems of marine meteorology and the knowledge required to attack them, since little attention is paid to these subjects in either text or diagrams. The treatment is often factual and in parts the method of presentation, together with the list of questions at the end of some chapters, seems designed to instruct the reader parrot-wise, rather than to direct him into courses of intelligent reasoning and profitable enquiry. An almost complete lack of references is both an omission and a handicap. For a book thus titled, too much emphasis is laid on the synoptic meteorology of land surfaces and of the upper air, also the treatment is rather conventional. Synoptic maps in the Appendices all are related to the United States mainland. In an analysis of the situation portrayed by one of these synoptic charts appears the bland statement ' On account of the lack of observations over the Atlantic Ocean the size of the area of bad weather cannot be determined accurately . . .' The large number of voluntary observing ships which nowadays regularly provide quite a reasonable network of observations in the relevant portion of the Atlantic will scarcely find this statement encouraging.

The chief fault of the book seems to lie in the fact that it attempts to cover so many varied aspects of meteorology without any attempt at specialization. At the same time subjects of the greatest importance for the mariner, such as climatology and oceanography, are dismissed in a few pages. The incidence of events at sea, such as fog development and dispersal, normal and anomalous radar propagation, ice accumulation on ships in high latitudes, besides the meteorological aspect of the care of cargo aboard a ship, all of which vitally affect navigation, and many other aspects of maritime activities in peace and war, receive scant if any attention in this volume.

In addition to the recurring trouble of faulty emphasis there is also a large number of loosely worded, obscure or inaccurate statements scattered through the book. When discussing the aneroid barometer, the authors refer to ' certain types of minor errors ' and make no reference to the necessity in practice for frequent comparisons with a mercury barometer. In describing pressure tendency the need to take account of the ship's own movement is neglected. Chinook winds are wrongly explained. Although in discussing upper-air observations radiosonde is mentioned, no description of radio-wind technique is included. A short account of the stratosphere makes no mention even of the existence of a negative or small lapse rate above the tropopause. Blocking action is badly defined and inadequately featured from a synoptic standpoint. A description of jetstreams, besides being superficial, assigns to them a significance which is hardly justified in this context.

From the mariner's viewpoint an even more serious and vital omission occurs in a description of the signs indicating the approach or development of a tropical cyclone, where no mention is made of the large diurnal variation of pressure found in the tropics, and the mariner is not shown how the magnitude of a pressure fall due to an approaching and developing tropical cyclone can be distinguished from a normal pressure fall due to diurnal variation of pressure in the region. The description of the creation of a tropical cyclone (p. 179) is most inaccurate and over-simplified and is contradicted on p. 180 by the sixth factor quoted therein as being responsible for their formation. Many other repetitions, contradictions, and inaccuracies were noticed, suggesting that a second edition of the book should be preceded by drastic editing. In particular, the reference to the International Meteorological Organization on p. 41 is surprising in a book claiming to be up-to-date.

On the credit side there are some excellent diagrams and photographs. In particular those of sea state reach a wide audience for the first time, and although in some cases they fail to give a realistic impression of the actual sea surface, they should nevertheless form a valuable contribution to this aspect of marine meteorology. Minor blemishes in the diagrams include some anomalous pressure gradients in equatorial regions (Fig. 6.1) and a very out-of-date diagram (Fig. 7.27) showing the weather of a ' typical anticyclone.' The inclusion of full details of weather reports, codes and analyses, costs of synoptic charts and so forth, while it has an advantage for readers at the present time, will tend to ' date ' the book all too soon. A series of synoptic charts of upper winds over the United States given in the Appendix has a relevance to the text which is far from obvious. A lack of detailed climatological information regarding many areas of the world is to some extent offset by the material included in a useful glossary, besides some colloquial terms. A satisfactory index could have been usefully extended to include the diagrams, and in a book with so many cross references, numbered paragraphs would find a place.

In compiling this book semi-officially the authors had an advantage over those responsible for any similar publication which is officially sponsored, and in certain respects they have exploited their situation. Certainly many of the diagrams have the merits of clarity and simplicity, and will make a direct appeal to the audience for whom they are intended; while those portions of the book, for instance, which deal with the importance of weather for strategic operations during the Second World War, and with meteorological aspects of the problems posed by radioactive fall-outs associated with hydrogen bomb explosions, enhance the book's value at the present time.

R.F.M.H. and C.E.N.F.

Theoretische Hydromechanik. By N. J. Kotschin, I. A. Kibel and N. W. Rose. Band II, translated from Russian into German by K. Krienes. Berlin (Akademie-Verlag), 1955. Pp. viii, 569; 156 Figs; 11 Tables. 48 DM.

Throughout my reading of this volume I have found myself comparing the text with corresponding books written in English and the volume is probably best judged in this light. The book serves as a text-book in the universities and technical schools of the U.S.S.R., and although it will interest all students of hydrodynamics it is probably of greatest use to students of aeronautics.

The book has three main divisions : 240 pages are devoted to compressible flow (Kibel), 260 pages mainly to incompressible viscous flow (Kotschin) and, it is pleasing to note, 50 pages on the elements of turbulence theory (Kibel).

The first chapter on compressible flow deals with the fundamental gas equations and in particular with the general equations governing the propagation of discontinuities. The second section deals with steady two-dimensional gas flows in which a clear exposition is given of the applications of characteristic theory in supersonic flow, dealing particularly with the aerodynamic problem of the force exerted on a thin two-dimensional profile, and also with the two-dimensional Tschaplygin theory and its development by Khristianovich for subsonic flow. The third section deals again with the aerodynamic problem, this time in the case of axially symmetric (subsonic and supersonic) flow past a solid of revolution. The only unsteady flow discussed is the well-known one-dimensional flow investigated by Riemann and Hadamard and the important omission in this division from the standpoint of the student of aeronautics is the flutter problem.

The second division dealing with viscous incompressible theory is most comprehensive in its content and most detailed in its treatment. In this particular field many of the notable advances have been made in this country and much of this appears in detail with adequate references. A brief section here relates to the boundary layer problem in viscous compressible flow and another is devoted to flows in which the viscosity is vanishingly small.

The third division is most welcome since it presents students with the now well-known Tollmien-Heisenberg-Lin theory of stability for two-dimensional flows, with the stability of Couette flow due to G. I. Taylor, with a discussion of the equations governing turbulent flow in liquids and with a brief historical survey of turbulence theories and some illustrative examples.

My notice concerning the first volume of this book was rather lukewarm but this second volume is one which can certainly be commended not only for the selection but also for the presentation of its material, and I feel that this could well serve as the basis for third-year university courses to fluid-motion specialists.

T.V.D.

Theoretical Hydrodynamics. By L. M. Milne-Thomson. London (Macmillan), 3rd Edition, 1955. Pp. xxi, 632; Figs.; Tables. 60s.

The first two editions of this book have found it a place as a standard text for undergraduates on those branches of hydrodynamics that can be treated by the use of a velocity potential. It is elegant and reliable, a bit fussy in its emphasis of vector methods, but an excellent reference book for the examinee. The third edition contains no substantial alterations, and it is more appropriate to review it not as a new book but as a re-issue of an old book in a new age. For hydrodynamics used to be thought of as a classical subject developed in the age of Kelvin, Rayleigh and Lamb, applied mathematicians being more attracted nowadays towards cosmical and nuclear subjects.

But however much his thoughts may turn to the majestic and the minute, man still has to deal with real fluids in his everyday life and they are a fit subject for his greatest philosophical skills. Professor Milne-Thomson says, and we would all agree, that the theory of ideal fluids was given a new lease of life by its explanation of the lift on an aerofoil; but that lease has been given a greater extension by the theory of the boundary layer, which he dismisses as follows : ' The theory of the boundary layer serves as a useful guide to experimental work and gives a qualitative description of viscous motion near a boundary, but its applications have so far been of a tentative and empirical character.' If by applications we mean the provision of exercises for undergraduates, then boundary-layer theory fails; but it has given a new understanding to fluid motion, and has presented applied mathematicians with real problems to subdue, that is problems requiring new philosophical methods.

Likewise turbulence is disposed of : ' The motions of our ideal inviscid fluid will always be supposed to be of the former (streamline) character. An exact mathematical treatment of turbulent motion has not yet been achieved.' Not yet ! Such a treatment is impossible on any sensible definition of turbulence; is it therefore never to be considered as part of theoretical hydrodynamics ?

Hydrodynamics has been revitalised in the last two decades simply because it has begun to deal with the exciting problems of boundary layers, turbulence, and motions with vorticity; and this has been as evident in meteorology as in any branch of the subject. The complaint therefore is that the many students who will turn to this book for guidance through their examinations will make no contact with hydrodynamics as it is today in the postgraduate world. There is plenty of inspiration for the dilettante, dabbling in the complex potential of flow past elliptic cylinders and the like.

In his preface the author remarks upon his extensive use of diagrams to bring a ' real ' subject to life. There are two criticisms to make here. One is that the drawings are not accurate. The sine curves have discontinuous curvature, the paraboloids with spherical caps are uniformly horrible, streamlines past a sphere run part of the way parallel with the surface. The draughtsman, who clearly did not know what it was he was drawing, has shaded his solids of revolution (which any student can understand with no difficulty from a plane section) as if this were the most elementary G.I. handbook; but there are very few flow patterns illustrated : this is the second complaint. The supreme example is in the section on ' the flow past a cylindrical log.' Here we see a perspective view of a rolling pin on a pastry board but no streamlines. Perhaps it does not matter, for no real fluid could be made to flow along the calculated streamlines.

This last example is fairly typical of the book in which application of the mathematical method is carried out with great elegance, but in which the reality of the problem seems irrelevant. Another good example of this is the flow spiralling outwards from a combined source and vortex – a practical impossibility. This could be, but is not, brought to life as the well known bath-plug experiment simply by reversing the flow. The mathematics would be unaltered. One has the impression that water and hydrodynamics are inhabitants of different worlds.

The last chapter on supersonic flow, which was not in the first edition, might be described as a concession to reality and is a very useful introduction to the method of characteristics. But it is not mentioned that a shockwave is always created by the steepening of a large-amplitude soundwave, nor that it can travel faster than sound.

These points are mentioned not because elegant mathematics for its own sake is not worth while; on the contrary it is delightful. It can, however, be combined with an adequate mention of the roots of the subject in order to encourage and inspire natural philosophers as well as pure mathematicians. This book does not tell the student what hydrodynamics is about : if that information and inspiration can be found elsewhere, he will find the book an example of pleasing elegance (diagrams excepted), and very helpful to him in his exercises.

If the study of ' theoretical hydrodynamics ' is justified by its explanation of the lift of an aerofoil how much more is it justified by all the other thousand and one realities which are its lifeblood today !

R.S.S.

Bibliography on land and water utilization and conservation in Europe. By C. H. Edelman and B. E. P. Eeuwens. Rome (Food and Agriculture Organization of the United Nations), 1955. Pp. vii, 347. Np.

This bibliography of the major subjects dealing with land and water use in Europe owes its production to the activities of a sub-committee on Land and Water Use of the European Committee on Agriculture. The titles listed represent about 50 per cent of the material collected and many smaller papers were omitted. It is grouped under the main headings : General and agricultural geography; Maps; Climate; Soils; Vege-

tation; Hydrology and hydrography; Land reclamation-Drainage-Irrigation-consolidation of fragmented holdings; Soil erosion and soil conservation. It comprises about 4,000 references relating to the various countries in Europe. The volume should be a most valuable and comprehensive reference for all engaged on work in this wide subject.

R.S.R.

Elements of ecology. By George L. Clarke. New York (John Wiley), 1954. Pp. xiv, 534; 201 Figs.; 22 Tables. 60s.

Dr. Clarke, Marine Biologist at the Woods Hole Oceanographic Institution, has provided a splendid collection of examples of the interactions of organism and environment, based on wide experience and wide reading. His own interest in marine ecology is dominant, but even if all this were removed the rest could stand as a first-class survey of terrestrial ecology. The amount of material handled is great and varied, but sensible arrangement and good writing have produced a book to be read with ease, pleasure and profit.

The formal meteorology is slight, but the organisms that steer their weather experiences into equations, tables and maps should find much to interest and amuse in discovering what temperature, wind, radiation and other weather elements mean to those organisms that haven't reached the stage of summarizing their experiences in the proceedings of a learned society. The author assumes that the reader needs no explanation of several complex phenomena in physical chemistry (e.g., osmosis), and it is occasionally necessary to sort out a confusion of force and pressure, of weight and density. In places enjoyable reading breaks into a chuckle. What visions of a new Beaufort scale there are in : ' In Texas Panhandle it is said that on a normal day a man can expectorate a mile and a quarter.' How accidental is ' Diurnal fluctuation in temperature is very much damped in the aquatic environment ' ?

Not least of its virtues, the book will provide excellent illustrations – verbal and pictorial – of the role of weather in the world of nature, and it offers a wide range of micro-meteorological problems challenging the expert to translate facts into physics.

H.L.P.

Proceedings of the Bournemouth Conference 1955. National Smoke Abatement Society, 1956. Pp. 131; Tables and Illustrations. 12s. 6d.

The National Smoke Abatement Society is now very well known. Perhaps its most important function is to bring together the people working on atmospheric pollution reform, to exchange information on this many-sided problem. For those who were unable to attend the 1955 conference at Bournemouth most of the information and much of its atmosphere is captured in this well-produced report. Highly significant parts were played by Mr. Alfred Robens and Mr. Gerald Nabarro, and by many representatives of Local Authorities, architects, engineers, chemists and members of the Beaver Committee.

A.R.M.

Papers in marine biology and oceanography. Supplement to Vol. 3, *Deep sea research.* Dr. Mary Sears (Ed.). London (Pergamon Press), 1955. Pp. xx, 498; Figs.; Tables; Plates. $10.50.

This supplement has been prepared to celebrate the 25th anniversary of the Woods Hole Oceanographic Institution and in honour of Dr. H. B. Bigelow, who has been intimately associated with it since its foundation. It contains a bibliography of Dr. Bigelow's writings and 48 papers covering a wide range in marine biology and oceanography. The contributions are from 113 authors, who were students or associated with him at the Institution. This volume will form a valuable reference for oceanographers, fishery biologists and other research workers in allied fields. It is handsomely produced and its publication was aided by a grant from the Woods Hole Oceanographic Institution.

R.S.R.

Obituary Notices

Dr. J. PATTERSON, O.B.E., M.A., LL.D., F.R.S.C.

John Patterson was born of Scottish parents in Oxford County, Ontario, on 3 January 1872, one of a family of thirteen children. He was brought up on his parents' farm and after receiving his early education at the local primary school, two and a half miles from his home, he attended the collegiate institutes at Ingersoll and at Woodstock, Ontario. He subsequently taught in a public school for five years to obtain the means to pursue his studies, intending to become an engineer. He entered the School of Practical Science, now the Faculty of Applied Science and Engineering, at the University of Toronto in 1896, and graduated in 1899 in the three-year course in civil engineering. Finding the courses not enough to occupy his energies, he also registered in the Faculty of Arts, taking the honours course in mathematics and physics, along with his engineering studies. During his fourth college year he was able to devote his whole time to his Arts course so that, when he graduated, he won the Gold Medal in Physics and also the 1851 Exhibition Science Research Scholarship, being the first winner of this scholarship from Toronto. This enabled Patterson to carry on post-graduate study for two years at the Cavendish Laboratory, Cambridge University, under Sir Joseph J. Thomson.

In 1903 Patterson became Professor of Physics in the University of Allahabad, India, and in 1905 was appointed to one of the newly created posts of Imperial Meteorologist to the Government of India. During this appointment he was actively interested in warnings of cyclones and in the public weather forecasts, which were issued only once per day. The great Indian earthquake of 1905 took place the first morning after Patterson took up his meteorological duties at Simla, and led to the establishment there of a seismograph which he set up and attended.

He returned to his homeland to accept the newly created position of Meteorological Physicist in the Meteorological Service of Canada. During World War I, at the request of the British Admiralty, he was loaned to them to take part in a survey being made in Canada for a source for helium, and towards the end of the war he succeeded in perfecting a process for the separation of helium at a cost of about $50 per thousand cubic feet. By this time, however, the United States had come into the war and helium was produced in great quantity from the Texas gas wells so that, although the Canadian project had met with some success, it was not continued.

After World War I Patterson returned to the Meteorological Service, continuing his earlier investigations on the upper atmosphere. He modified the Dines meteorograph for Canadian use to reduce the risk of destruction of the record when the instrument struck the ground. He also developed the Canadian pilot-balloon programme and devised a simple procedure for computing the velocity of the upper winds from the flight observations. Patterson was mainly interested in wind and pressure instruments. He will perhaps be best known for the development of the three-cup anemometer which he originated and which, he was able to show, had definite superiority over the four-cup type. He also made intensive studies, along with others, on improving the anemometer by beading the edge of the cup and changing the shape from a hemisphere to a cone with a straight lip. Later, he developed a satisfactory electromagnetic anemograph for recording wind speed and direction.

Until 35 years ago, 75 per cent of the mercury barometers shipped to Canada arrived broken, with the metal parts contaminated by mercury. Shortly after World War I, Dr. Patterson decided to design a barometer, for manufacture in Canada, which would combine the advantages of a Kew Barometer with the portability of the Fortin barometer. This instrument, now known as the Patterson barometer, can be shipped to any weather station in Canada with little risk of breaking. He also developed an almost automatic procedure for filling barometer tubes with pure mercury, a method still in use today, in which the boiling of mercury is eliminated and there is no danger of breaking the barometer tube.

He was appointed Assistant Director of the Canadian Meteorological Service in 1924 and Director in 1929, the title being changed to Controller in 1936. As Director, or Controller, he was responsible for modernizing the Service and encouraging the introduction of the latest developments in all branches of meteorology. He organized meteorological services for Trans-Canada Air Lines and for the Canadian side of trans-Atlantic aviation, and, with the outbreak of the Second World War, for the Royal Canadian Air Force. He retired from his official position on 1 December 1946.

For over thirty years he worked faithfully on the executive of the Royal Society of Canada and the Royal Canadian Institute. He was elected into Fellowship of the Royal Meteorological Society on 17 November 1920 and to Honorary Membership on 19 January 1941. He greatly enjoyed his connection with both the Royal Meteorological Society and the American Meteorological Society, and was President of the latter in 1930-1932.

Patterson's wisdom was greatly respected in international meteorology and he represented Canada in numerous conferences and committees. He was President of the Commonwealth Conference on Meteorology in London, England, in 1935, and was President of the W.M.O. Technical Commission on Instruments and Methods of Observation, 1946-1953. From 1940 until 1947 he was honorary professor of meteorology at the University of Toronto.

A very happy event in India for Dr. Patterson was his marriage to Margaret Norris, M.D., born at Staffa, Perth County, Ontario, who had gone out as a Medical Missionary and was in charge of the Seward Hospital of the American Presbyterian Mission at Allahabad. For 50 years Mrs. Patterson created an ideal home life for her husband.

Patterson belonged to the Presbyterian Church (later the United Church of Canada) and took an active part in the Congregation to which he belonged, carrying out his duties as an office-holder (elder). His religious faith pervaded his daily life and gave him a quiet tenacity of purpose that carried him triumphantly through frequent periods of delay and disappointment. He was a non-smoker and was strongly opposed to the drinking of alcohol. For a few years prior to the Second World War, he was a member of a Curling Club and devoted the one evening a week he felt he could spare from his office work to playing in a team.

Patterson obtained his greatest pleasure in his work. He loved designing instruments and carrying out his administrative duties. Fortunately, he had a strong constitution so that he was able to work for ten or twelve hours a day for many years without taking sick leave or the authorized annual holiday. After his retirement from official duties, he continued to come to the office daily and carried on his writing and instrument work until six weeks before his death, which occurred on 22 February 1956.

Dr. Patterson made a notable contribution to the advancement of meteorology in Canada and abroad by his untiring energy, his sound judgment and the integrity of his character.

Andrew Thomson.

Professor WILFRED SMITH

Wilfred Smith, John Rankin Professor of Geography in the University of Liverpool, died suddenly at the university on 27 September 1955 at the age of 52. Throughout the whole of his academic life he was associated with Liverpool University. After graduating with first-class honours in Geography in 1924, he became Tutor in the Department, and then successively Assistant Lecturer (1928), Lecturer (1931), Senior Lecturer (1945) and Professor (1950). Furthermore, since 1953 he had been Dean of the Faculty of Arts, in 1954 he was the President of the Institute of British Geographers, and just before his death he had been elected to the Royal Society's National Committee for Geography. It was in the sphere of economic geography that, by scholarly research and numerous publications, he had become such an eminent authority, and his *An economic geography of Great Britain* was published in 1949. He also maintained a life-long interest in meteorology, joining the Royal Meteorological Society in 1924, encouraging the study of meteorology and climatology within the Department of Geography, and publishing climatic studies of China (1934), the Fylde (1936) and Merseyside (1946).

S.G.

Mr. H. W. APPERLEY

Mr. Harry William Apperley, who died in Morecambe Hospital at the age of about 77 on 13 December 1955, after a long illness, was born at Belthorn near Blackburn. In 1895 he entered the service of the National Telephone Company at Blackburn, whence his duties later took him successively to Barrow-in-Furness and Manchester. He was transferred to the Post Office Engineer's department at Lancaster in 1912, and in 1915 returned as an executive officer to Blackburn, where he remained until his retirement in 1938. Mr. Apperley had been an enthusiastic student of meteorology since boyhood. He kept regular weather records during most of his life and became known to a wide public through his reputedly very accurate forecasts for the northern English counties, published in various local newspapers. Summaries of his climatological observations appeared in the *Northern Daily Telegraph* from 1919 until the outbreak of war in 1939. He was elected a Fellow of this Society in 1930, under the sponsorship of the late Sir Richard Gregory, F.R.S., at that time President.

E.L.H.

The Reverend I. C. ROBERTS

The Reverend Isaac Charles Roberts, one of the Society's oldest Fellows, died at Prestatyn on 19 March 1956 as the result of an accident. He was in his 92nd year. Born at Cyffylliog, Vale of Clwyd, Mr. Roberts spent his early youth as a farmer and pupil teacher in his native village. After a period at Denbigh Grammar School and training at Bala College he entered the Welsh Presbyterian ministry in 1887. The longest and last of his successive pastorates, none of which took him very far from his birthplace, was at Gwernymynydd, near Mold, where he stayed for 20 years. For some time he was chairman of the board of guardians for the Holywell Rural Council. He retired in 1926 and went to live at Prestatyn. Keen interest in the weather had led Mr. Roberts to join this Society nine years earlier, and now the science became something more than his hobby. From 1930 to 1946 he and his wife were the official meteorological observers under the municipal authority at Prestatyn. Later he published a summary of the records in booklet form. He had long given special attention to the abundant weather lore of Wales. At the age of about 87 he contributed an entertaining article on this subject to the issue of the magazine *Weather* for November 1951. Many of the items therein were translated by himself into English out of the original tongue, and a number of the sayings cited are not to be found in Richard Inwards's great collection. Mr. Roberts was latterly the oldest surviving Welsh Presbyterian minister.

E.L.H.

THE NAPIER SHAW MEMORIAL PRIZE

Announcement of second competition

The Royal Meteorological Society announces the second competition for the Napier Shaw Prize. The competition will be open to anyone without restriction of nationality. The prize of £100 is offered on this occasion for an original essay on **'Climatic variation.'**

> The essay may deal with climatic variation on any time scale or scales. A critical analysis and appraisal of the evidence is sought concerning a large part of the earth's surface and preferably on a global basis. The use, where suitable, of modern statistical theory in the interpretation of the evidence will be welcome. Theoretical explanation in geophysical terms may also be offered.

The essay, which must be in English, may include material which the candidate has already published. Essays may be submitted jointly in the names of more than one author.

Candidates are required to submit three copies of their essay, which must be printed or typewritten. In the latter case, only one side of the paper should be used, double spacing should be employed and two of the copies may be good carbon copies. Unless the essay is printed, any mathematics should be written in carefully by hand. Although the scale of the essay will be considerably greater than that of an ordinary paper, candidates are recommended to consult, for general advice, the pamphlet, 'Preparation of papers for the Quarterly Journal.' This may be obtained, free of charge, on application to the Assistant Secretary.

It is required that any essay for which a prize is awarded shall be offered to the Society for publication.

All entries for the competition should be addressed to the Assistant Secretary, Royal Meteorological Society, 49 Cromwell Road, London, S.W.7 and marked clearly 'Napier Shaw Prize.'

Essays for the competition must be received at the Society's Offices *not later* than 1 January 1959.

ROYAL METEOROLOGICAL SOCIETY
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To mark the completion of ten years' publication of *Weather*, the Royal Meteorological Society offers a prize of 40 guineas for an essay of about 5,000 words in English under the title :

'The place of meteorology in liberal education'

The competition is open to anyone. Entries must be received at the Society by 31 December 1956. The prize-winning essay will become the property of the Society and may be printed in *Weather* or any other suitable journal.

Additional prizes may be awarded if entries of high enough standard are received but competitors may state that they wish to enter for first prize only; for such additional prizes particular consideration may be given to entries from persons under 25 years of age.

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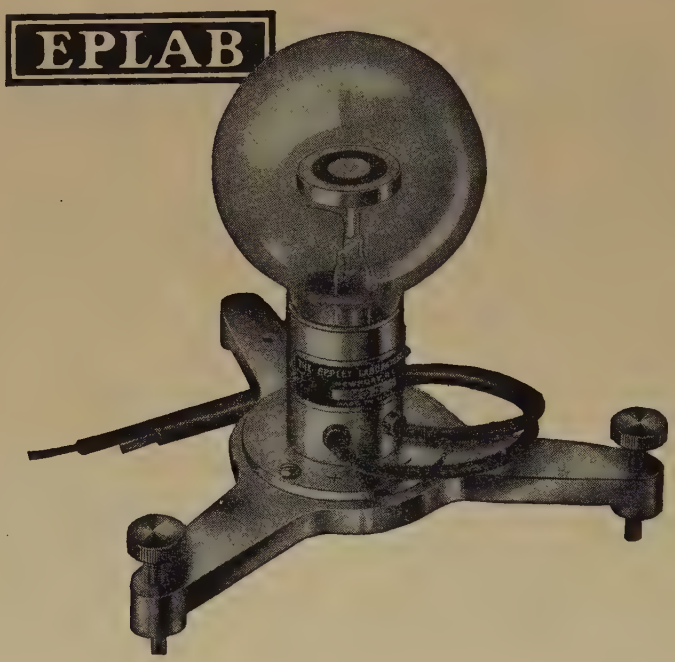

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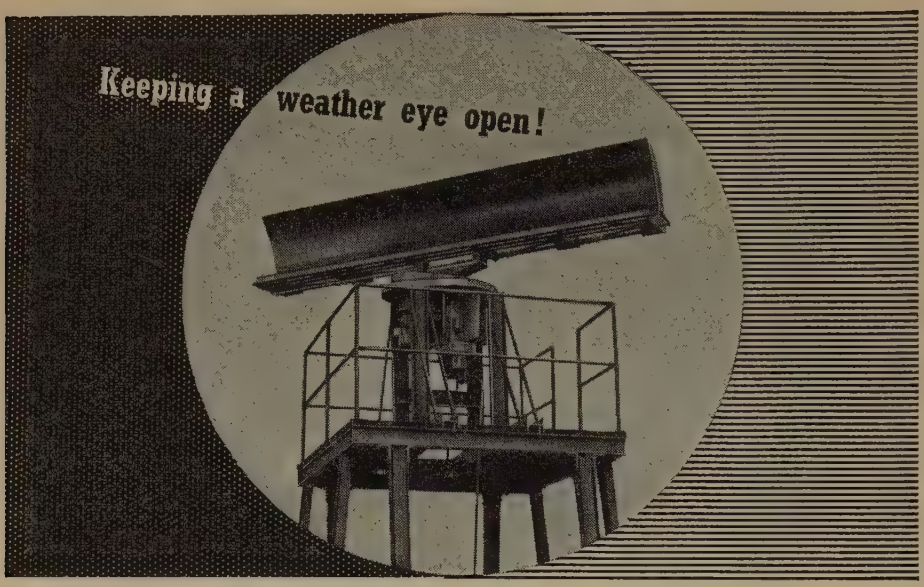

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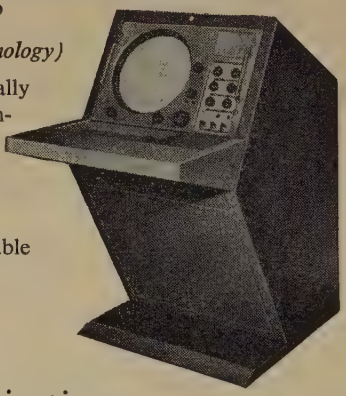

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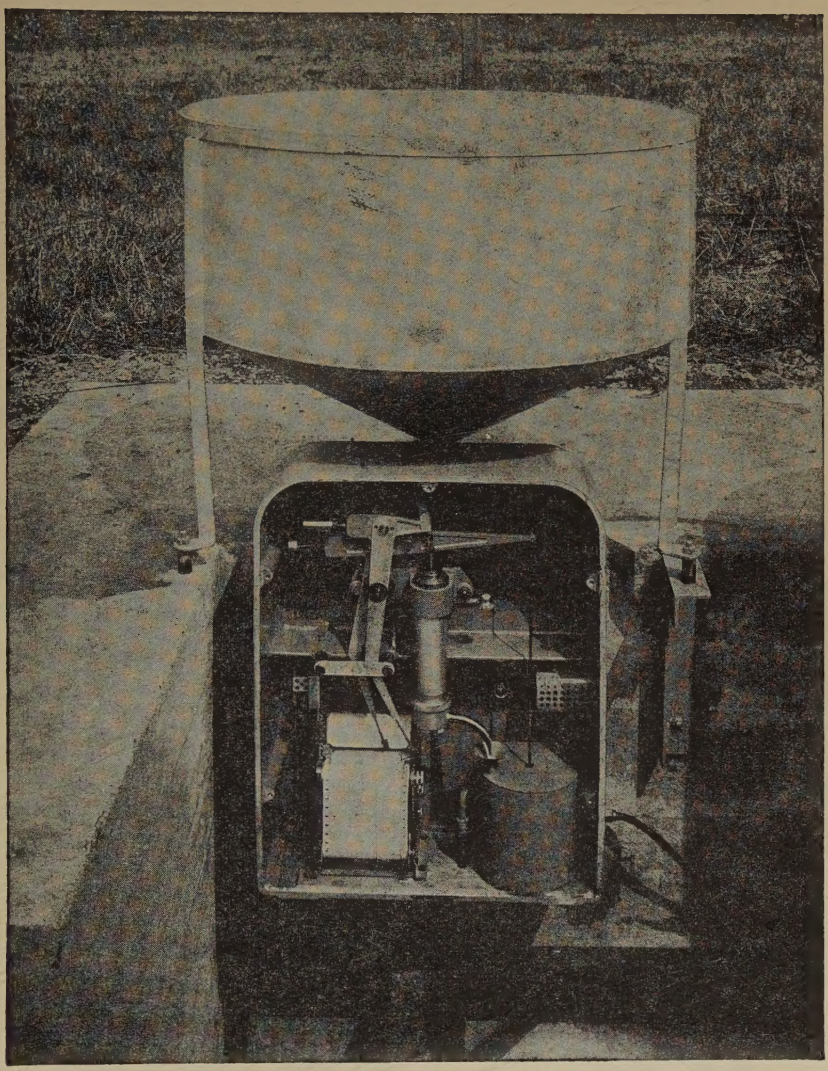

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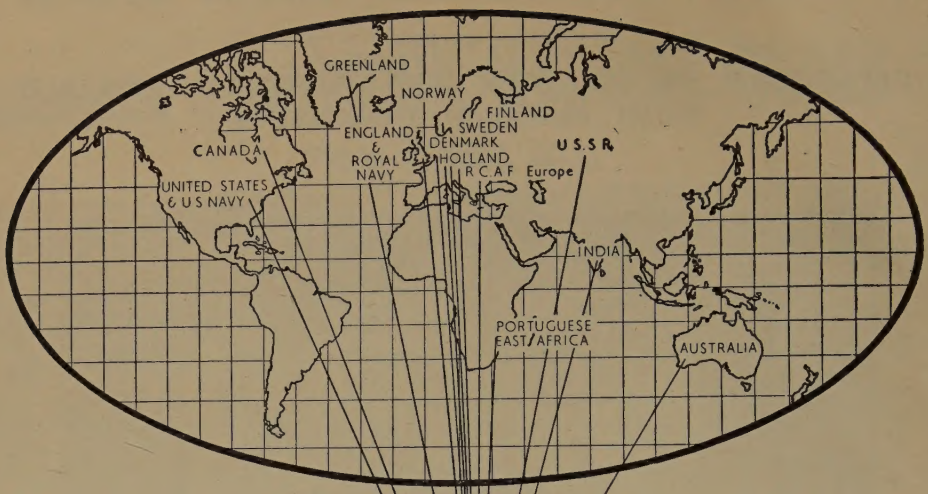

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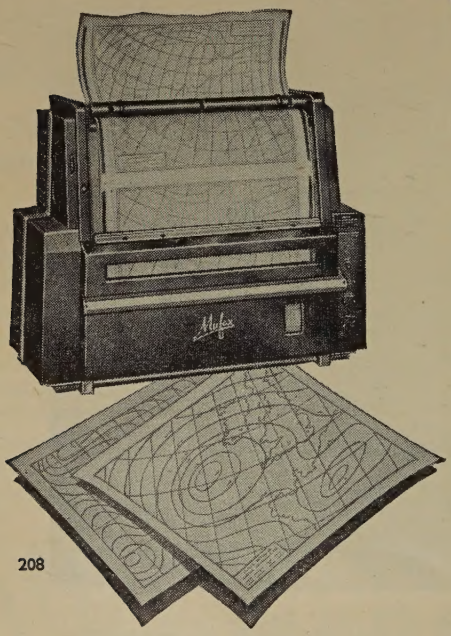

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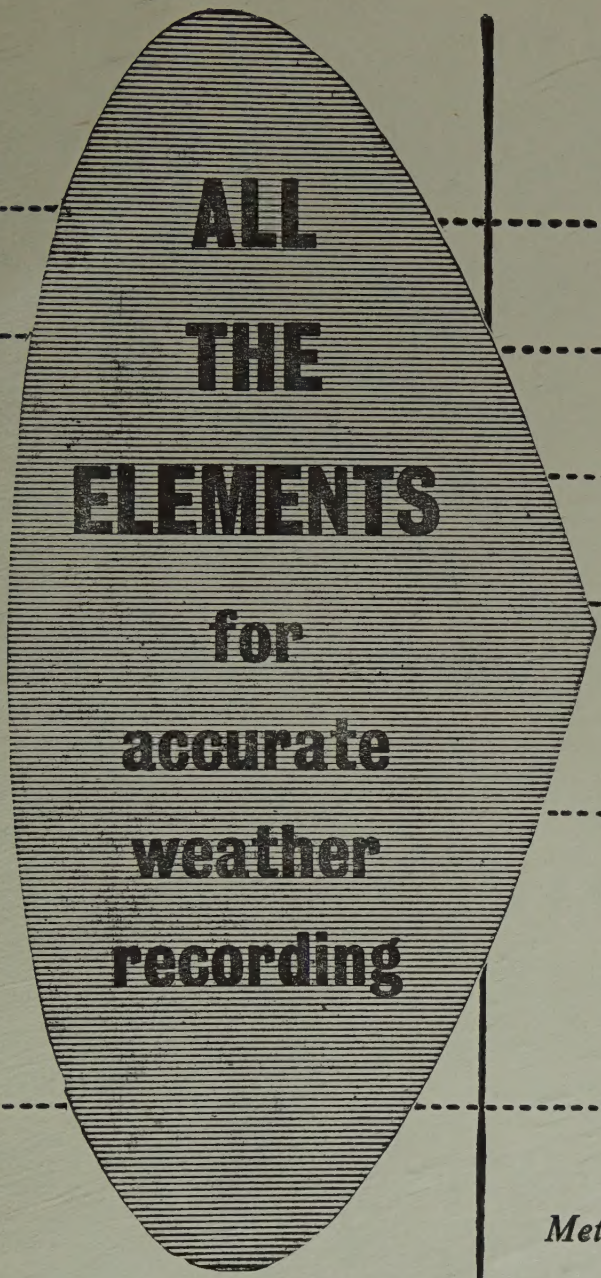

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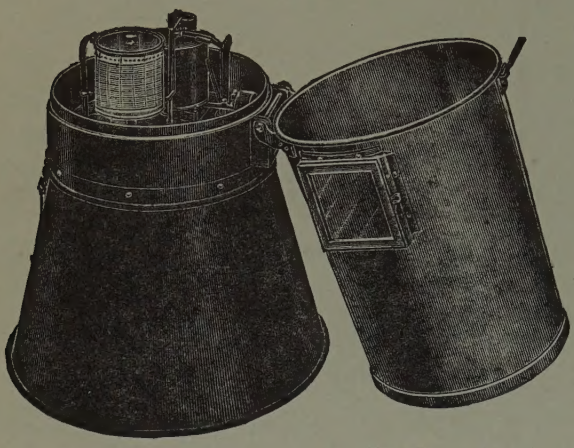

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